THE SNOW ENVIRONMENT

OF THE CRAIGIEBURN RANGE

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T. D. PROWSE

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a_k	albedo of snow surface to shortwave radiation (0.0 - 1.0)
d_s	snow depth (m)
$\bar{d}_s$	mean monthly snow depth (m)
e_a	near surface vapour pressure (mb)
e_a'	adjusted sea level vapour pressure (mb)
e_s	snow surface vapour pressure (mb)
e_ν	saturation vapour pressure (mb)
e_ν'	saturation vapour pressure adjusted to sea level (mb)
$e_{\nu w}$	saturation vapour pressure at wet bulb temperature (mb)
g	acceleration due to gravity (m s^{-1})
k	von Karman's constant
q	specific moisture of air (kg kg^{-1})
t	time (d)
u	wind velocity (m s^{-1})
z	depth or height (m or mm)
z_a	wind speed measurement height (m)
z_b	humidity or air temperature measurement height (m)
z_o	surface roughness parameter (m)
z'	adjusted instrument heights (m)
C_a	specific heat of air (J kg^{-1})
C_i	specific heat of ice (J kg^{-1})
C_w	specific heat of water (J kg^{-1})
D_E	transfer coefficient of latent heat
D_H	transfer coefficient of sensible heat
D_M	transfer coefficient of momentum
E_a	elevation of upper site (m)

E_b	elevation of reference site (m)
F_P	elevation of freezing level at time of maximum precipitation (m)
F_T	elevation of freezing level at time of minimum air temperature (m)
H	average horizon angle from the zenith (degrees)
$K\downarrow$	global radiation (W m^{-2})
K^*	net shortwave radiation (W m^{-2})
L_f	latent heat of fusion (J kg^{-1})
L_v	latent heat of vapourization (J kg^{-1})
$L\downarrow$	atmospheric radiation (W m^{-2})
$L_o\downarrow$	atmospheric radiation from clear skies (W m^{-2})
$L\uparrow$	terrestrial radiation (W m^{-2})
L^*	effective radiation (W m^{-2})
P	precipitation intensity (mm s^{-1})
$\bar{P}$	total storm precipitation (mm)
P_a	near surface barometric pressure (mb)
P_v	snow surface vapour pressure (mb)
Q_E	latent heat flow (W m^{-2})
Q_G	ground heat flow (W m^{-2})
Q_H	sensible heat flow (W m^{-2})
Q_M	heat available for melt (W m^{-2})
Q_P	precipitation heat flow (W m^{-2})
$Q_{\dot{P}}$	latent precipitation heat flow (W m^{-2})
$Q_{P'}$	combined precipitation heat flow (W m^{-2})
Q^*	radiation heat flow (W m^{-2})
Q_θ	heat deficit heat flow (W m^{-2})
R	ram resistance (kg)
Ri	bulk Richardson number
T_a	air temperature ($^\circ$C or K)
T_a'	adjusted sea level air temperature ($^\circ$C or K)

$\bar{T}_a$	mean monthly air temperature (^{o}C or K)
T_d	dry bulb air temperature (^{o}C or K)
T_g	temperature of ground (^{o}C or K)
T_p	temperature of precipitation (^{o}C or K)
T_s	temperature of snow (^{o}C or K)
T_w	wet bulb air temperature (^{o}C or K)
T_B	long term air temperature at reference site (^{o}C or K)
T_x	daily maximum air temperature (^{o}C or K)
Vf	view factor (0.0 - 1.0)
W	total daily windrun (km d^{-1})
α	empirical constant (10.0)
γ	ratio of molecular weights between air and moisture (0.622)
ε_a	emissivity of atmosphere (0.0 - 1.0)
ε_s	emissivity of snow (0.0 - 1.0)
ζ	heat deficit (J m^{-3})
λ	ground thermal conductivity (W m^{-2} K^{-1})
ρ_a	density of air (kg m^{-3})
ρ_s	density of snow (kg m^{-3})
ρ_w	density of water (kg m^{-3})
σ	Stefan-Boltzman constant (5.6697 x 10^{-8} W m^{-2} K^{-4})
ϕ	fraction of sky obscured by cloud (0.0 - 1.0)
ψ	coefficient of recrystallization
Ω	coefficient accounting for cloud height and temperature
ℓ	long term mean lapse rate (^{o}C or K m^{-1} or km^{-1})
$(\ )_s$	stable atmospheric conditions
$(\ )_u$	unstable atmospheric conditions

ABSTRACT

The snow environment of the Craigieburn Range is addressed under three major themes: snowfall, snow metamorphism and snow melt. A six year record of alpine climate and three years of snow structure analysis, which form the most extensive and reliable record of alpine snow conditions yet available in New Zealand, act as the data base.

Overall, the results indicate that the Craigieburn Range snowcover is most typical of intermontane or coastal-transition regions. Snow storms and snowpack structure exhibit some characteristics typical of both maritime and continental climates. Snowfalls are usually small magnitude, low intensity events, although on average one extreme event occurs each year. Snow storms often contain periods of rain, and are characterized by temperatures which rarely fall below -6^{o}C. In the absence of strong winds, new snow densities are similar to those for continental regions.

Changes in the density and structure of deposited snow are rapid because of the relatively warm nature of the snowpack. Equi-temperature metamorphism produces density increases of up to 50% per day and is further hastened by frequent melt and rain periods which occur during the main winter period. Indirect climatic evidence suggests temperature gradient metamorphism is unlikely. Field investigations, however, demonstrated that depth hoar crystals occur throughout the snowpack and frequently develop in conjunction with ice crusts, another dominant

feature of the snow stratigraphy.

Sensible heat flow is the major source of heat in both winter and spring melt periods. On a daily basis, net radiation is of secondary importance, although during the daylight hours, solar radiation is usually the largest heat supply. The greatest total heat transfer to the snowpack occurs on days with rainfall. For these days, precipitation heat flow is small but the latent heat released by condensation exceeds that from net radiation.

Days of high sensible heat transfer, characterized by high winds and warm air temperatures, are related to the occurrence of north-westerly winds. The amount of evaporation during many of these days is equal to high values measured overseas.

The implications of an intermontane classification for the Craigieburn Range is also reviewed relative to two practical considerations, avalanche forecasting and snowmelt hydrology.

CHAPTER I

INTRODUCTION

1.1 IMPORTANCE OF SNOW TO NEW ZEALAND

Snowcover forms an important part of the physical environment in the world's cold regions and as a result has been the focus of extensive and diverse research by a variety of academic disciplines. Snowcover in New Zealand is primarily limited to the alpine zones but the total area above the estimated winter snow line is considerable, being in excess of 38,000 km^2 or 13% of the total land area. Despite the considerable size of New Zealand affected by snowcover, prior to the 1960s, snow had received only minimal scientific investigation. Since that time, a body of literature has accumulated dealing with a variety of aspects of snow. However, much of the research has been conducted in an unconsolidated fashion and the overall picture of the New Zealand snow environment remains unclear.

The belated and desultory nature of snow research in New Zealand may generally be ascribed to two reasons. Firstly, almost the entire population of New Zealand permanently resides below 700 m in relatively snow-free lowland areas. Very few public roads extend above even 1000 m (Simpson-Housley and Fitzharris 1979). Because of this geographical relationship snow has played a relatively minor role in man's activities. Similarly, limited opportunity has existed for the observation-documentation of of the snow environment as regularly occurs in continental climates where snow descends to quite low levels or in other mountainous countries where a high population concentration is found within the alpine zones.

Second, there is no central well-funded agency or body in New Zealand, as exists in many other countries, whose primary directive is the study of snow science and technology. Most research has been conducted by a variety of agencies in which research interests only fringe on the snow environment.

During the last two decades, the importance of snow to New Zealanders has greatly increased primarily because of the demand placed on the alpine snowcover in areas of resource development and recreation. Increasing demands upon water resources has created the need to better understand the natural processes which generate streamflow. In 1969 the Technical Subcommittee on Snow (TSS) of the New Zealand National Committee for the International Hydrological Decade outlined the need for more research concerning the New Zealand snowcover in terms of total water storage and the benefits of forecasting snowmelt runoff for flood mitigation, irrigation and hydro-electricity generation. The TSS tentatively identified snow regions in which snow courses could be established to obtain representative snow data and listed a number of broad areas for further research in snowmelt forecasting.

The importance of snowcover has also been recognized by the New Zealand tourist industry, especially in the areas of ski-field development and planning. Increases in overseas visitors and the growth of the local skiing fraternity has created an ever-increasing demand for more ski-fields (Pearce 1977). Owens and O'Loughlin (1979) estimate that existing ski-fields have experienced a 300 - 400% increase in patronage over the last decade. Related to this growth has been a greater demand for information regarding the planning, development and operation of ski-fields (Owens and Prowse 1980). The type of information required has ranged from aspects of snow hydrology to assist in road construction and streamflow routing, to data concerning the overall reliability of snowcover.

An additional point of interest for the alpine recreation industry has been the hazard due to snow avalanching. An assessment of the hazard by La Chapelle (1979) revealed that a number of New Zealand ski-fields have a high avalanche risk. Prowse et al. (1981) calculated that there has been over a 300% increase in the number of deaths due to avalanching between the 1960s and 1970s, although a majority of the fatalities were climbing rather than skiing related.

Snow avalanching and large snowfalls have also caused severe disruption to travel within New Zealand and on occasion have effectively isolated entire communities for long periods. Snow avalanching on the Milford Highway during 1980 closed the route for over six weeks and cost the tourist industry alone more than $17,000 per week (Prowse et al. 1981). A number of key highway passes are also regularly closed due to snowfall. Expensive road maintenance is the only method by which road links between the west and east coast of the South Island may be maintained during large parts of the winter season.

The large impact on transportation, communication, forests, agriculture and buildings from heavy snowfalls occurring outside the winter season and/or in lowland areas has also been well documented [Hughes (1969, 1974); Tomlinson (1970); Paul (1980)].

In summary, as greater demands have been placed on the snowcover resource and more use made of the alpine areas, the need for a greater understanding of snow science and technology in New Zealand has increased. The next section details the current body of knowledge concerning New Zealand snowcover.

1.2 SCIENTIFIC LITERATURE

The first scientific work on snow in New Zealand appeared in 1962 with Heine's (1962) description of surface snow densities on Mt Ruapehu. Subsequent investigations have primarily appeared under four major themes: snow accumulation, hydrology, avalanches and meteorology. Important aspects of many of the studies dealing with the above topics appear in the main text of this report but the following outlines the general foci.

Early work by Archer (1970) on the Ben Ohau Mountains, and by Gillies (1964) and Chinn (1969) in the Fraser and Waitaki catchments, provided cursory examinations of snow distribution and accumulation. More extensive work, describing in particular relationships of accumulation to elevation, ground cover and wind flow have been provided by Fitzharris (1972, 1976), Harrison (1978), Morris and O'Loughlin (1965); O'Loughlin (1969a), Weir (1979) and Weir and Owens (1981). The distribution of snow in the context of glacier mass balances has also been analyzed by Anderton (1975, 1976a, b); Anderton and Chinn (1978); Kells and Thompson (1970); and Thompson and Kells (1973) on the Ivory, Tasman, and Whakapapanui glaciers.

The role of seasonal snowcover in the production of runoff was first considered in detail by O'Loughlin (1969b). A number of researchers [Anderton (1974); Grimmond (1980); Owens and Prowse (1980); Prowse (1980b)] have since attempted to model snowmelt discharge primarily based on predictive models developed overseas. Other studies have focused more on the relevant importance of individual components of the energy balance in producing snowmelt [Harding (1972); Anderton and Chinn (1978); Prowse (1980a, b)]. Fitzharris et al. (1981) has also discussed the role of meltwater in flood production.

A large number of avalanche reports containing primarily subjective assessments of snow and associated weather conditions are contained within various New Zealand alpine journals (Owens et al. in prep.). The first scientific account of snow avalanching did not appear until 1976 in Fitzharris' (1976) description of an avalanche event in the Mt Cook area. McNulty and Fitzharris (1980) subsequently described another event in the Craigieburn Mountains and Weir and Owens (1981) have described the significance of snow structure for avalanche occurrence on

Mt Hutt. McGregor (1980) has attempted the first statistical prediction of avalanches in New Zealand, while Conway (1977), McIntosh (1979), Fitzharris and Owens (1980) and Waters (1980) have mapped known avalanche tracks.

Information concerning the meteorology of snowfalls in New Zealand is primarily confined to an early summary by Kidson (1932), concerning snowfall frequency and annual variation, and a number of reports dealing with either long term trends or specific conditions of 'exceptional' snowfalls [Burrows (1976b); Donaldson (1974); Hughes (1969, 1974); Neale and Thompson (1977); Tomlinson (1970)]. These reports have largely dealt with lowland snowfalls and only O'Loughlin (1969a) has considered snow storms in alpine areas.

1.3 RATIONALE - OBJECTIVES

In contrast to the importance of snowcover in New Zealand, the quantity of existing scientific literature must be considered somewhat limited. To exemplify this point many of the research goals outlined by the Technical Subcommittee on Snow remain uninvestigated. Only marginal progress has been made in the further definition of snow regions or even the simple quantification of the size of the snow resource. This lack of research is apparent by the small number of snow courses currently operating in New Zealand. A conservative estimate of the ratio of snow courses to area of snowcover in New Zealand is 1 to 4000 km^2. Within the San Juan Mountains of the United States the ratio is only 1 to 268 km^2 (Caine 1975) and is even lower at 1 to 77 km^2 within the Snowy Mountains of Australia (Snowy Mountains Hydro-electric Authority 1970). This simple comparison indicates an apparent deficiency in even the most elementary snow data. More specific topics of necessary future research have been outlined for a variety of disciplines including: hydrology (Fitzharris 1979), meteorology (Neale and Thompson 1977) and avalanche dynamics [La Chapelle (1979); Dingwall (1976, 1980)].

Information dealing with specific snow-related problems can be obtained from individual scientific investigations or, in many cases, from the more general body of existing knowledge dealing with snow science and technology. The laws governing the forms and processes of snow are universal and hence a great deal of information is interchangeable from one location to another. However, care must be taken of the extent to which 'imported' information is used to the exclusion of original research for any given location. As propounded by Colbeck (1979, 1) "too often we have generalized the results of local research to the rest of the world. Information which is relevant to one region may not be applicable to other regions even though the same laws of nature apply to snowcover anywhere." According to Colbeck (1979) there must be a recognition of the wide variety of conditions under which snow is (a) deposited, (b) undergoes metamorphism, (c) melts, and (d) runs off. The identification of these parameters for any location is an important first step to maximizing the benefits of inter-regional exchanges of snow science and technology. The main purpose of this study was to increase the knowledge of these basic parameters for a particular snow region, the Craigieburn Mountains. Therefore, the following three major objectives were defined:

(a) to describe the characteristics of snowfall emphasizing the temporal distribution of precipitation types, magnitude and intensity, and the nature of snow deposition.

(b) to evaluate the metamorphic processes and crystal structures which dominate the stratigraphy of snow once it has been deposited.

(c) to calculate and compare the heat flows which are important to snowmelt and to isolate the conditions under which the most rapid melt occurs.

Unfortunately, an investigation of the above mentioned fourth aspect of snowcover, the conditions under which snow runs off, was found to be beyond the resources of this study. Some work was conducted on this aspect but the findings remain tentative and require further investigation.

A more general goal of this study was to compare and contrast many of the results to overseas research and in the process develop a broad climatic definition of the Craigieburn Mountain snowcover.

1.4 PRESENTATION FORMAT

Six chapters comprise the remainder of this report.

The first of these describes the selected study area in terms of location, physical setting and climate, and outlines the methods of data collection. Measurement accuracy is also briefly considered and the chapter concludes with a consideration of spatial and temporal representativeness.

Chapters three, four and six present the major findings of this study with each chapter focusing on one of the three major objectives. Because of the relatively distinct nature of these chapters, separate sections describing the relevant literature, background theory, sampling methodology and analysis are contained within each. The background for chapter six was so extensive that a separate chapter (five) was devoted to its introduction.

Chapter three describes the characteristics of snowfall in the Craigieburn Range with respect to temporal variations in precipitation type, intensity and magnitude, freezing levels, and synoptic patterns.

Chapter four investigates the development of snowpack stratigraphy based on both indirect meteorological and direct field evidence. Special attention is paid to the dominant metamorphic processes and their effect on snow structure.

The theory of snowmelt forms the major part of chapter five. Each component of the energy balance is considered with respect to snowmelt generation and equations are described by which each heat flow is calculated in chapter six.

Chapter six focuses on the calculation of heat flows to the snowpack and their relative importance in producing snowmelt during both spring and winter months. The significance of snow evaporation is also evaluated.

The final chapter reviews the major findings of each chapter and assesses the significance of some of the results to the points of interest concerning snowcover outlined in section 1.1. Areas of future research are also considered.

CHAPTER II

STUDY AREA
SITE DESCRIPTION AND INSTRUMENTATION

2.1 GENERAL LOCATION

The general study area lies in the centre (43^{o} 10' S, 171^{o} 40' E) of the Craigieburn Range, a 26 km chain of mountains, 20 km to the east of the main divide of the Southern Alps of New Zealand (figure 2.1). The range lies in a predominantly north-south direction approximately equidistant (100 km) from the Tasman Sea and Pacific Ocean. To the west lies a series of mountains culminating in the main divide of the Southern Alps and to the east, the Castle Hill Basin, Torlesse Range and finally the Canterbury Plains.

2.2 PHYSICAL DESCRIPTION

2.2.1 Geology-Topography

The Craigieburn Range contains a number of peaks over 2000 m but the average elevation is approximately 1800 m. The geology and topography is typical of much of the eroded mountain lands east of the Southern Alps, characterized by rounded ridge tops and slope angles of 30 to 40^{o} above 1200 m. Bedrock outcrops frequently appear above 1500 m, but the lower elevations are overlain by thick talus, solifluction deposits and morainic or alluvial gravels.

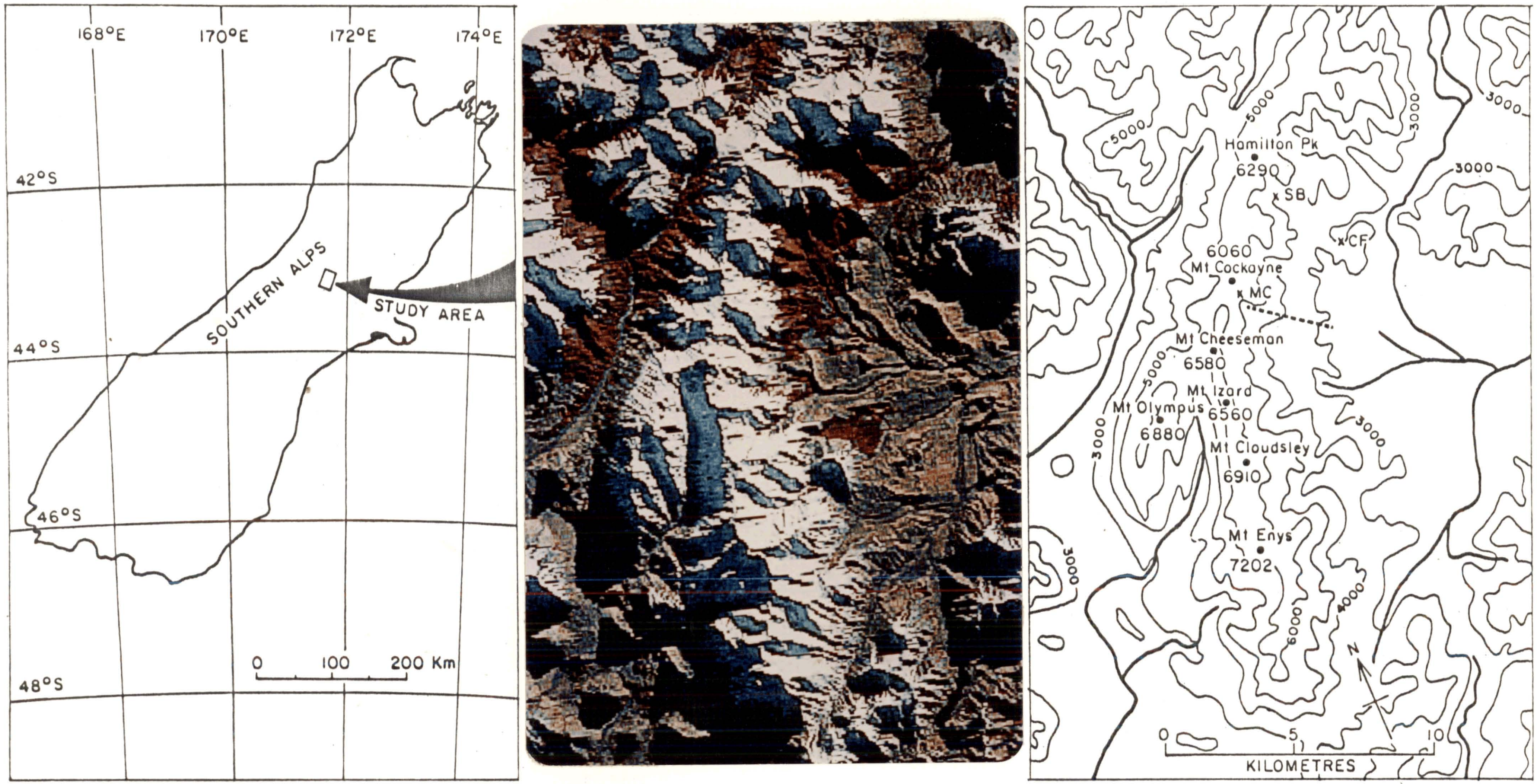


Figure 2.1 STUDY SITE LOCATION. The centre photo is a LANDSAT image of the Craigieburn Range enhanced to detail snowcover. SB, MC and CF refer to the three climate stations discussed in the text. The dashed lines indicate the location of the thermistor profile.

Most traces of the earlier U-shaped valleys, cut in the Pleistocene and Recent times, have been destroyed by fluvial action or hidden beneath talus deposits. Similarly, former cirque basins, dominating the upper elevations of the eastern side of the range have been noticeably modified (Chinn 1975).

The basement rock is comprised of thick beds of strongly indurated sandstone and siltstone of the Triassic-Jurassic age. Beds are commonly tilted and faulted with numerous shatter belts (Gage 1980).

2.2.2 Soils

The soils of the area are classed as Bealey within the forests and Spenser in the grasslands, both having a skeletal, steepland, yellow brown earth appearance. The A horizon is very shallow owing to the low erosion resistance of the top and subsoils.

2.2.3 Vegetation

The lower slopes are covered in part by mountain beech forest (*Nothofagus solandri* var. *cliffortioides*) with canopy heights of about 20 m decreasing to only 10 m at timberline (1370 m). In areas where snow drifts accumulate against the upper forest margins dense krummholz, 1 - 3 m high develops (Wardle 1965). Grasslands occupying the higher elevations are dominated by snow tussock (*Chionochloa pallens* and *C. macra*). The upper extent of grassland is

1850 m but much of the alpine area is covered by screes (McCracken 1980a). Both the forests and grasslands have been modified to varying degrees by fire and grazing.

2.3 CLIMATE

2.3.1 General

The Craigieburn Range lies perpendicular to the path of the prevailing westerlies at a distance of approximately 30 km inland of the point of peak precipitation on the South Island (McSaveney 1978). Much of the weather in the area follows a relatively simple pattern associated with the passage of successive anticyclones separated from each other by a meridional trough of low pressure (Hill 1961). The average time between high pressure systems has been estimated by de Lisle (1969) at approximately one week.

The standard pattern can be seen to begin with a trough travelling eastwards across the Tasman Sea accompanied by a steady build up of cloud cover and westerly winds. As the system continues eastward, rain normally falls along the Main Divide, especially if the wind is north-westerly. North-westerly precipitation frequently occurs over the Craigieburn Range but will only reach the Canterbury Plains if there is a disturbance in the normal airflow (Hill 1961). Beyond the eastward penetration of north-westerly precipitation the air is usually warm and dry, the effect comparable to a Chinook or Föhn wind (Lamb 1970). At the same time, an area of low pressure frequently forms in the lee zone of the mountains over the Canterbury Plains (Sevele 1969).

As the cold front within the trough passes over the country, the winds shift to a more southerly direction, temperatures decrease and precipitation may continue. In some cases the centre of depressions associated with the trough will pass directly over the country or tend to stall off the east coast. These situations can cause some of the most intense and persistent precipitation along the eastern and inland portions of the South Island. The trough is eventually replaced by another anticyclone preceded by cool southerly winds.

2.3.2 Mountain Climate Classification

The general state of mountain climatology is such that a clear definition does not exist for most of the alpine regions in New Zealand. Many of the early alpine climate descriptions have been based on indirect evidence, such as from vegetation [Wardle (1964); Zotov et al (1938)] and snowcover (Cockayne 1928) or, by the extrapolation-observation of meteorological data from low level elevations [Elder (1959); Garnier (1946); Zotov et al. (1938)]. In more recent years, many more mountain climate studies have been undertaken, primarily as part of ecological studies, but the number of permanent alpine climate stations has remained small in comparison to the extent of mountain lands in New Zealand. Figure 2.2 illustrates the locations of all climatological stations above 900 m currently listed in the

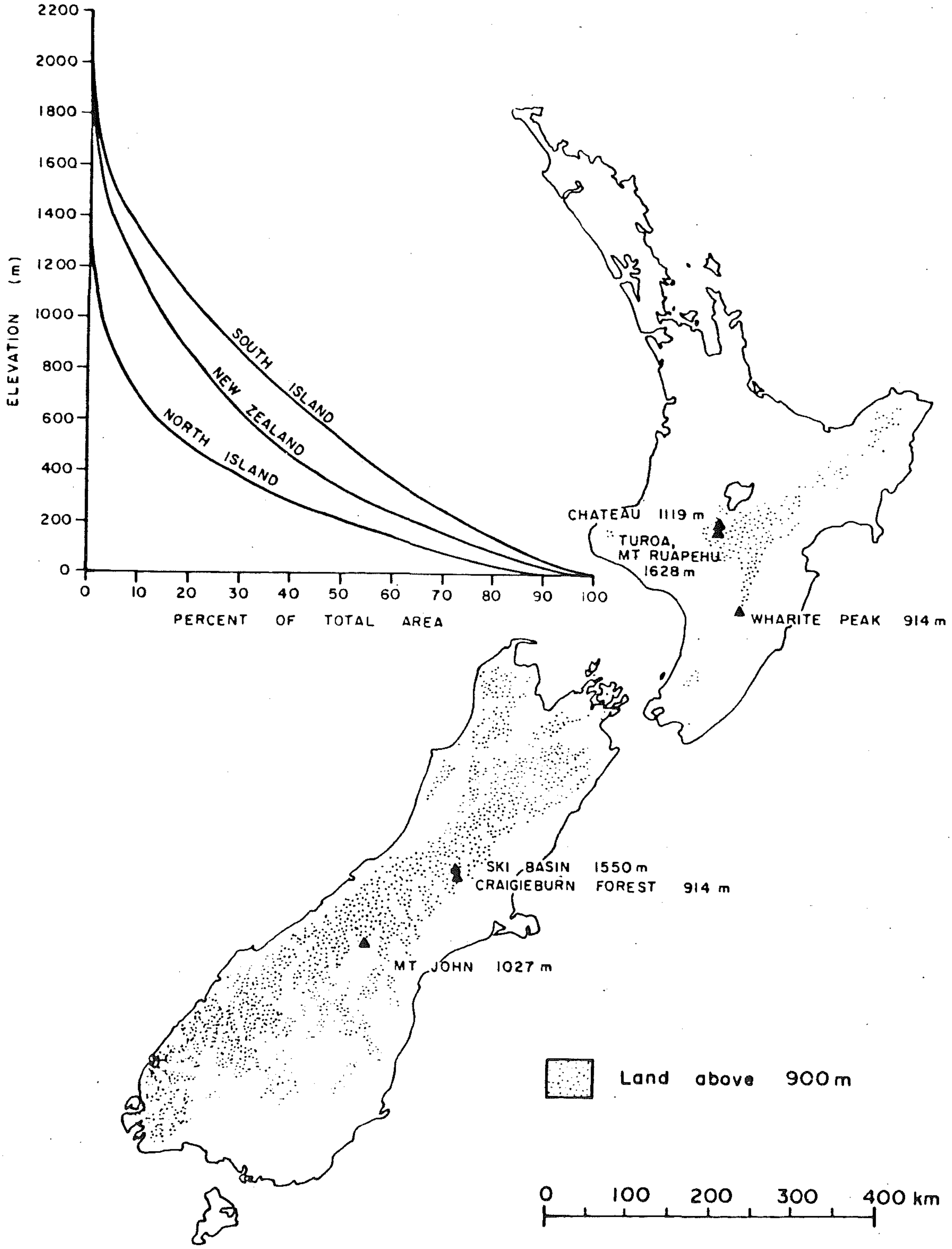


Figure 2.2 HYPSOGRAPHY AND ALPINE CLIMATE STATIONS OF NEW ZEALAND.

meteorological summaries of the New Zealand Meteorological Service (1980). Although 20% of the New Zealand land area is above 900 m the six stations noted on figure 2.2 represent only 2% of the total number of climate stations in New Zealand as reported in the above meteorological summary. A worse situation prevails in the South Island where three stations serve 29% of the total land area. Two of these stations are within the field area of this study.

Other alpine climate data have been collected at temporary stations which, in addition to regularly reported climate data, is the basis of Coulter's (1973) description of three contrasting categories of alpine climate for New Zealand:

(a) Areas near and west of the Main Divide (particularly south of Arthur's Pass) with high precipitation (2,500 - 8,000 mm yr^{-1}) and much cloud, fog and drizzle. Egmont, the Southern Ruahines and the Tararuas are classed as similar.

(b) The Central Otago Mountains with much less precipitation (1,000 - 1,800 mm yr^{-1}) but frequent fog and strong persistent winds. Low mean temperatures ranging from +3^{o}C in January-February to -6^{o}C in July.

(c) The eastern leeward ranges of Canterbury typified by the Craigieburn Range with more precipitation (1,500 - 2,500 mm yr^{-1}) but less fog, cloud cover and longer periods of drying winds. Similar range as (b) in mean 1500 m temperature but warmer, +10^{o}C in January to -1^{o}C in July. Inland Marlborough similar but drier. Kawekas also similar but wetter and milder.

The latter description is based on climate data collected primarily at the alpine station within the Craigieburn Range, which is described in more detail in the following section. Many more comparisons of the Craigieburn Range to overseas climates are included in this report and others are offered by McCracken (1980a).

2.3.3 Climate of the central Craigieburn Range

Climate studies in the Craigieburn Range began with the construction of a field station by the New Zealand Forest Service in 1959. The first published climate studies appeared six years later with a review of the overall climate by Morris (1965) and the results of some snowcover investigations by Morris and O'Loughlin (1965). Since that time a number of unpublished Forest Research Institute internal reports have detailed further studies, considering in particular: windflow (Rowe 1968), snowcover (O'Loughlin 1969a,b), precipitation (Rowe 1970) and general climate (McCracken 1980a).

The long term means of air temperature, precipitation, wind speed and radiation from two elevations (914 and 1550 m) in the Craigieburn Forest Park, which have been compiled by McCracken (1980a) are illustrated in figures 2.3 and 2.4.

In a study of precipitation between 1965 and 1968 Rowe (1970) calculated the mean annual precipitation as

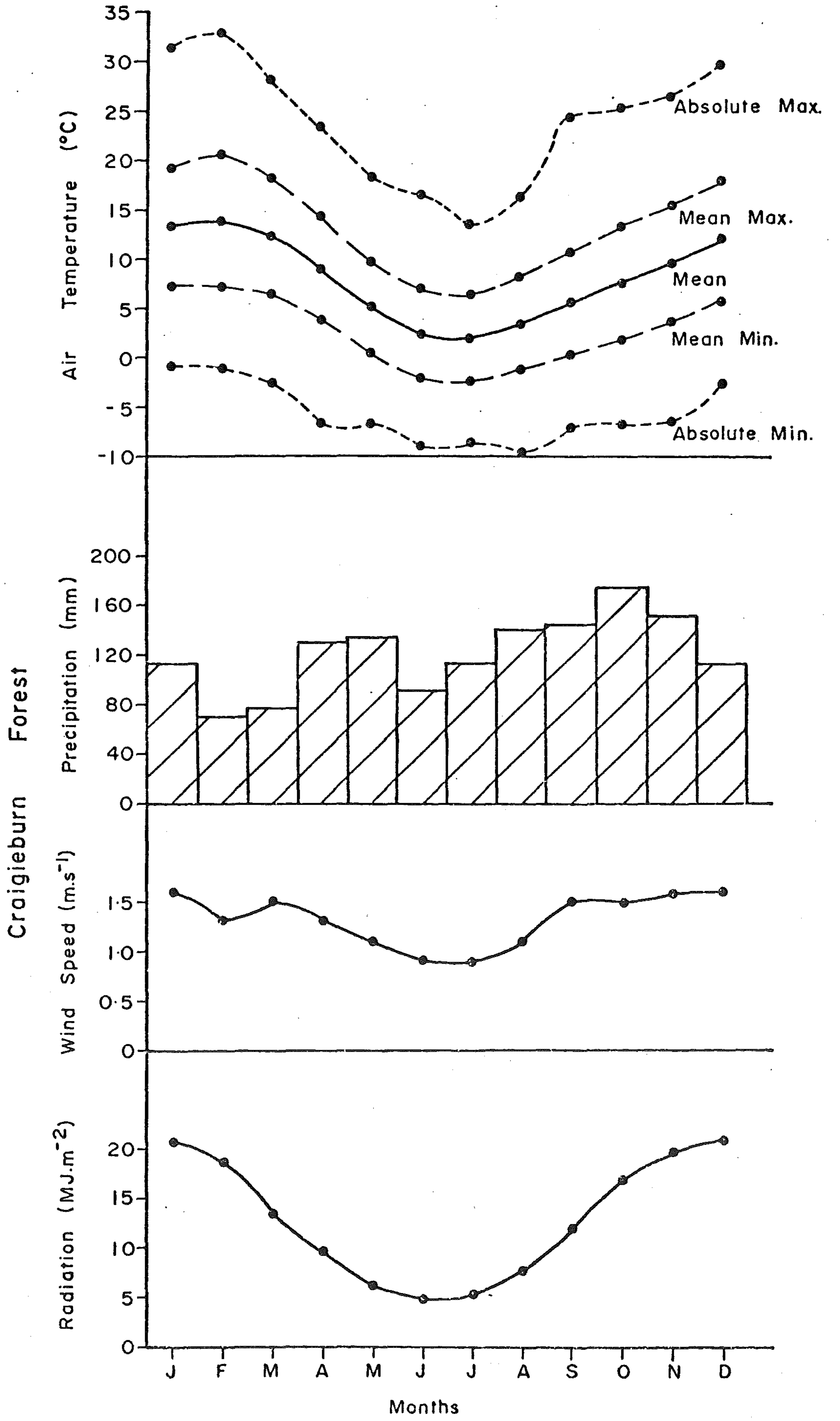


Figure 2.3 LONG TERM CLIMATE RECORD FOR CRAIGIEBURN FOREST CLIMATE STATION.

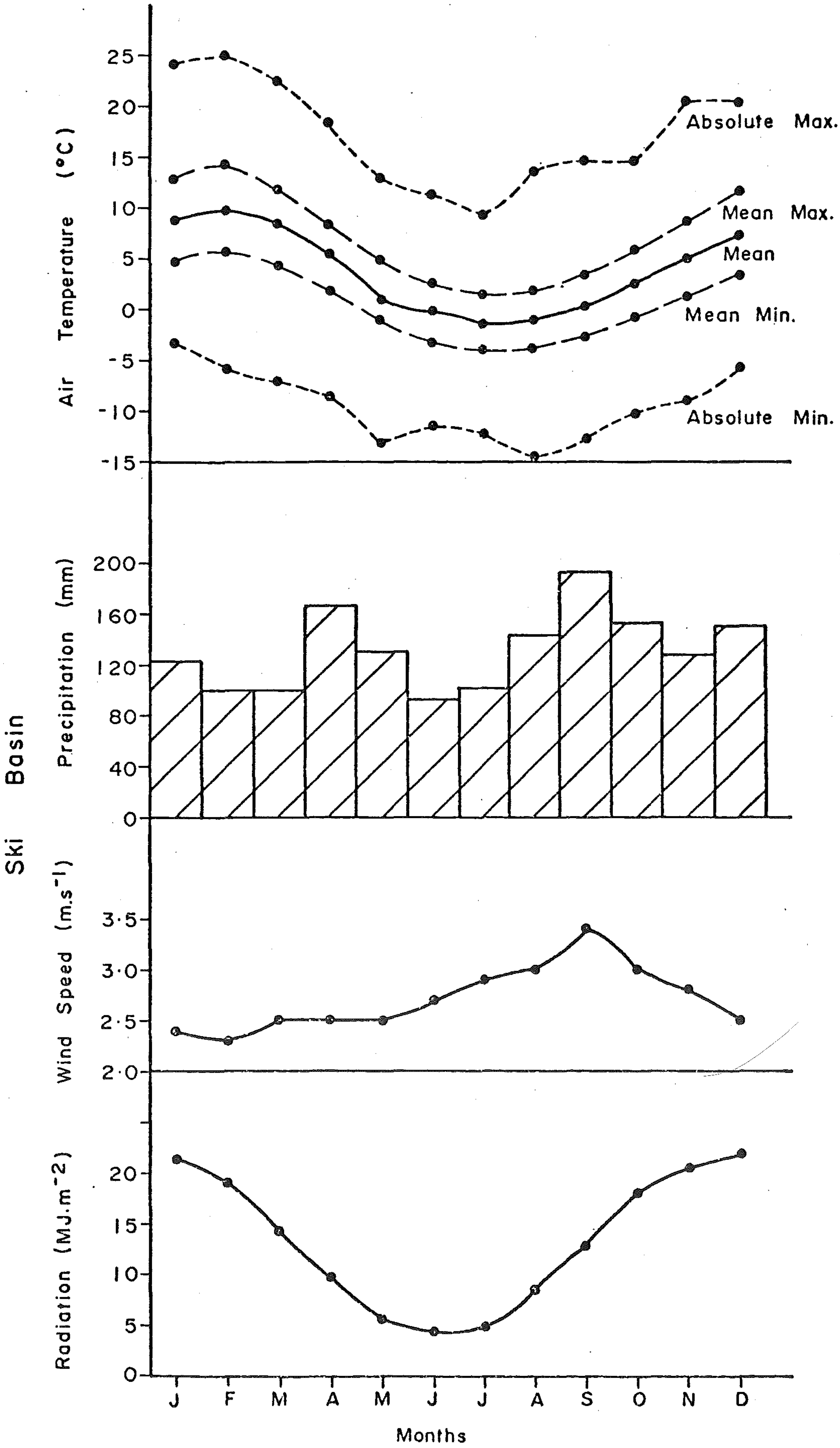


Figure 2.4 LONG TERM CLIMATE RECORD FOR SKI BASIN CLIMATE STATION.

1473 mm at 914 m increasing to 1778 mm at 1550 m. Over a fourteen year period, 1966-1979, McCracken (1980a) obtained a similar result for the lower elevation but a significantly less amount of 1586 mm for the higher site. Late winter to early spring appears to be the wettest period, perhaps because of an increase in north-westerly storms. Over 80% of total precipitation originates in airflows from the north-west to south-west quarter. Snow is likely to fall at any time of the year and according to Morris and O'Loughlin (1965) contributes over one third of the total precipitation above 1500 m but produces only negligible quantities below 900 m. The main snow accumulation occurs above treeline, usually beginning in May or June and reaching a maximum in September or October. The total duration of snowcover has been found by O'Loughlin (1969a) to vary from five to nine months.

At 1550 m, July is the coldest month with a long term mean temperature of -1.4^{o}C and February the warmest at 9.7^{o}C. The temperature range is not as great as found in continental climates (only 18.1^{o}C separates the absolute maximum and minimum temperatures) but diurnal fluctuations can be quite large, especially during a shift in wind direction from the north-west to the south.

At the lower elevation site (914 m) temperatures are warmer, the mean temperature difference between the two sites being 5.5^{o}C in January and 2.7^{o}C in June.

Global radiation varies from approximately 20 $MJ\,d^{-1}$ in the summer to less than a quarter of this value in mid-winter. Wind speeds are between about 1.0 and 1.5 $m\,s^{-1}$ at the sheltered low elevations but increase appreciably at the alpine location. In addition, there is a marked September maximum in the 1550 m mean wind speed (3.4 m s^{-1}) which coincides with the seasonal maxima for precipitation and may also be due to a concentration of north-westerly storms in the spring period.

2.4 INSTRUMENTATION AND DATA COLLECTION

2.4.1 General

A wide variety of instrumentation was required to satisfy the objectives of this study, outlined in chapter one. In particular, an extensive weather record was needed to evaluate the energy balance of the snowcover and to classify the nature of precipitation inputs. Although much of the snowpack structure data was manually collected in field surveys, some remote continuous recordings of temperature were also necessary.

A number of instruments were installed specifically for this snow research project but numerous problems were encountered in their operation, largely because of the harsh nature of the alpine environment. In all cases, these instruments had to be relocated at least once, before suitable long term recording sites were found. Since much of the data collected from these particular instruments were from a variety of locations, and contained large gaps due to instrument failure, a heavy reliance was placed on the existing climate recording network operated by the Forest

Research Institute.

The particular data records and methods of analysis employed in this study are detailed in the relevant chapters but the following briefly outlines the overall instrumentation network.

There are two climate stations located within the study area. The first station, Craigieburn Forest (CF), is located at 914 m in an exposed forest clearing on a 15^{o} slope facing a north-west direction. The second station, Ski Basin (SB), is approximately 5 km to the north-west of the CF site (figure 2.5) and is located at 1550 m on a 2^{o} slope at the base of a south-east facing cirque. Climate data were also collected at a number of elevations on a south-east facing slope of nearby Mt Cockayne (MC).

The instruments located at the various sites are outlined below according to the type of data collected.

2.4.2 Air Temperature

At both the CF and SB sites, Casella thermohygrographs were positioned in Stevenson meteorological screens at a height of 1.5 m above the ground surface (figure 2.6). Five other thermohygrographs were located on Mount Cockayne at 890, 1120, 1330, 1550 and 1680 m but the latter one was found too difficult to service and was subsequently relocated at a lower elevation.

Air temperatures were also obtained from an automatic temperature recording system at MC which was comprised of six cable connected field stations and a base station.

Figure 2.5 SKI BASIN CLIMATE STATION (1550 m)

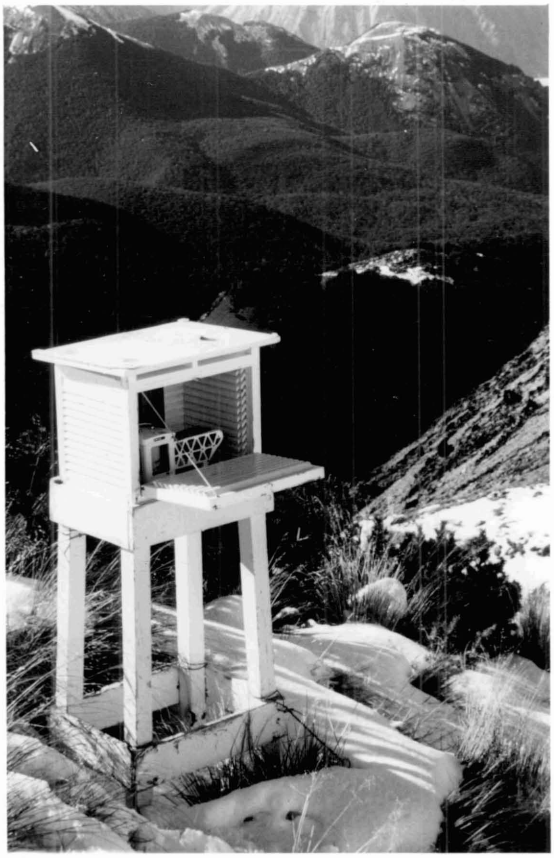


Figure 2.6 THERMOHYGROGRAPH LOCATED IN STEVENSON METEOROLOGICAL SCREEN, MT COCKAYNE.

Figure 2.7 THERMISTOR STATION, MT COCKAYNE.

Each field station contained nine thermistors, one at 1500 mm and two each at 300, 50, -50 and -150 mm from the ground surface (figures 2.7 and 2.8), recording either air, snow or ground temperatures on a fifteen minute cycle. McCracken (1980b) provides a complete description of the system.

2.4.3 Snow and Ground Temperatures

Snow temperatures were obtained from a Weather Measure distance thermograph which was first located at 1300 m in Camp Stream (figure 2.9) and subsequently moved to a higher elevation site (1680 m) on Mount Cockayne. The 300 and 50 mm thermistors, described above, also provided measurements of snowpack temperature whenever the snow was deep enough to completely cover the sensors. Ground temperatures were obtained from the -50 and -150 mm thermistors. Since only a shallow intermittent snowcover existed over many of the lower field stations only the record from the 1670 m station was employed in this study. Ground temperatures at a depth of 300 mm were also recorded by a distance thermograph at the SB climate station.

2.4.4 Precipitation

Belfort weighing bucket dual traverse precipitation gauges (200 mm orifice) were in operation at both the SB and CF sites. An antifreeze salt solution was used in the SB gauge to limit evaporation and increase the retention of snow. Prior to 1975 a variety of gauges were employed at

Figure 2.8 SURFACE THERMISTORS, +300 mm and +50 mm.

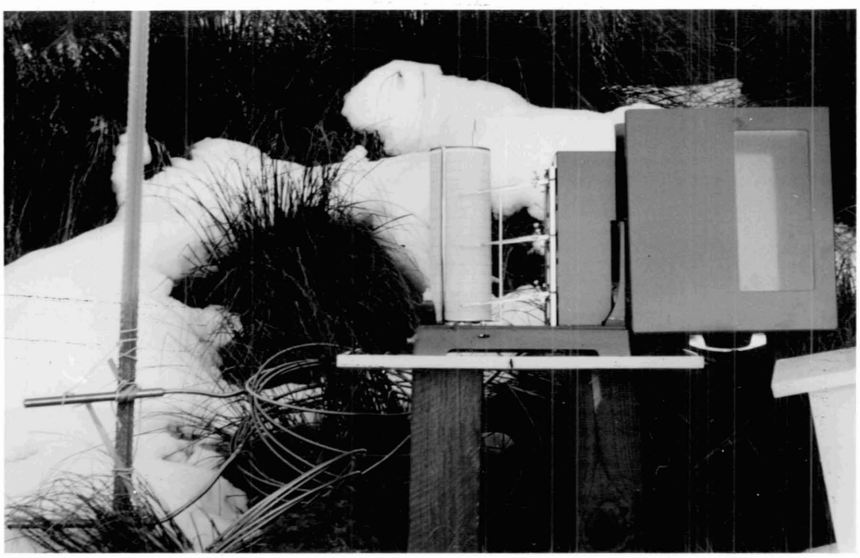


Figure 2.9 THREE POINT DISTANCE THERMOGRAPH

the two sites (Rowe 1970) and noticeable discrepancies existed in the total gauge catch between CF and SB. The installation of the Belfort gauge in March 1975 and the use of antifreeze has made the SB precipitation record the most reliable and accurate yet available for an alpine location in New Zealand. Although over the long term, a greater gauge catch has still been noted at the lower CF station, this author believes that in some cases the magnitude of the difference may be overestimated (example: arithmetic errors are suspected in the precipitation summary by Child (1978); 85% overestimation of the CF precipitation total from the manual rain gauge).

A Weather Measure propane heated snow gauge was installed at 1680 m on MC (figure 2.10) but because of severe drifting snow problems the gauge also had to be relocated at a lower elevation.

2.4.5 Windflow

A Lambrecht anemograph (instrument height = 6.1 m) measured windspeed and direction at the CF site while daily wind run was obtained from a totalizing cup anemometer (instrument height = 2.7 m) at the SB station. Another Lambrecht anemograph has since been installed on Mount Cockayne.

2.4.6 Radiation

Global radiation was measured by a Feuss bimetallic

Figure 2.10   MT COCKAYNE CLIMATE STATION (1680 m)

actinograph at CF and a similar instrument had been located at SB over the years 1969-1974. This six year concurrent record allowed an estimation of radiation receipts at SB from CF data over the period of this study.

2.4.7 Daily Climate Observations

Standard daily climate observations were made at the SB and CF sites by staff of the Forest Research Institute. Measurements of new and total snow depth were already a regular part of the daily observations, but during 1980, snow density profiles were also included in support of this study.

2.4.8 Upper Air Data

The nearest radiosonde station to the Craigieburn Range is at Christchurch, a distance of 80 km to the east. Observations by radiosonde are taken at least once a day from the Christchurch International Airport. Temperature, freezing level, pressure thickness and wind direction values were all extracted from the Christchurch records.

Information concerning the duration and nature of weather patterns were also derived from pressure charts. Surface pressure charts are available for New Zealand at a scale of $1:2 \times 10^7$ and $1:5 \times 10^6$ on a six hour interval while the upper level charts are only at the larger scale and a twelve hour interval. The standard thickness chart used by the New Zealand Meteorological Service is the 1000 - 500 mb.

2.4.9 Field Snow Surveys

The major part of the snow fieldwork involved the collection of snow pit data according to the methods outlined in Perla and Martinelli (1976). In particular, information was collected concerning snow density, depth, hardness, strength, grain type and size, temperature and free water content. Descriptions of the methods, equipment and classifications used in the snow pit observations are contained in appendix A.

During 1978 and 1979 snow pits and depth surveys were conducted over a wide range of elevations and aspects in order to establish a background concerning the variability in snowpack characteristics. In 1980, snow pit analysis was primarily conducted at weekly intervals at approximately 1800 m on both a shaded south-east facing slope and a sunny north-east facing slope of Mt Cockayne.

2.5 INSTRUMENT ACCURACY

2.5.1 General

A number of measurement problems are frequently associated with snow studies and many of these are magnified in the alpine environment. The following details some of these problems and outlines the ways in which most were accommodated. Problems associated with the use of data for particular analyses are discussed in the relevant chapters.

2.5.2 Mechanical Operation - Time Checks

The reliability of mechanically driven instruments in harsh environments, especially when affected by freeze-thaw cycles, is highly variable. A majority of the instruments described in previous sections rely on mechanical clock-drive mechanisms which produced, under the most severe weather conditions, recording time irregularities of up to four hours per day. Daily time checks were made of the thermohygrograph records at the SB and CF stations and these were used in the chart analysis to adjust any time variations to within an accuracy of approximately 0.3 hours.

2.5.3 Sensor Accuracy

The Casella bimetallic temperature sensors normally have a temperature error of approximately $\pm$ 1^{o}C. However, this error margin may markedly increase when the response of the sensor becomes damped from icing under conditions of high moisture levels and low air temperatures. Daily temperature checks made from glass thermometers at both climate stations were used whenever possible to correct the thermograph traces. In severe cases of instrument malfunction, the data were discarded and extrapolation performed from one of the other instruments.

The accuracy of the hygroscopic hair humidity sensors is estimated to be within 5% over the relative humidity range of 20 to 90%. Regular calibrations with the use of an Assman aspirated hygrometer were made in an attempt to retain the sensor response within this range but during sub-freezing conditions there are two additional sources of error. The first deals with the presence of ice on the hair sensors. In such conditions, vapour transfers between the sensors and the surrounding air is governed by vapour pressure over ice rather than water. In addition, water in the solid state can affect the mechanical response of the hair material. However, these problems are not believed to appreciably affect the results of this study because:

(a) the difference between saturation vapour pressure over ice versus water reaches a maximum of only 0.27 mb at approximately $-10^{\circ}C$ (pressure differences are negligible near the freezing point).

(b) a majority of the humidity data used in analysis were recorded during periods of above freezing air temperatures.

The second problem relates to the use of hair hygrometers at low ambient air temperatures where only small differences in moisture content produce large changes in relative humidity values. If the sensors are not continuously ventilated, Bergen (1968) estimates that in conditions of near freezing air temperatures, a time lag of up to an hour may result between the instrument and the surrounding air. In this study, most humidity values were averaged on a daily basis and hence the problem was considered minor.

Ventilation problems for the thermohygrographs also existed when the Stevenson screen shelters became blocked by blowing snow or accretions of rime. Although the screens and instruments were regularly cleared irregularities in the records from poor ventilation did exist especially during some intense storms. For these periods, extrapolation of temperature and humidity was made from other sites whenever possible or the data were discarded.

Large snow accumulations also presented problems in data recording because the instruments were not located on height-adjustable supports. During periods of large snow depth and minimal air turbulence, inversion conditions over the snow surface may have affected the instrument readings of temperature and humidity. However, because of the relatively windy nature of the environment strong thermal inversions were considered to be rare. In addition, whenever profiles of temperature and humidity were assumed, such as in the estimation of turbulent exchange, instrument heights above the snow rather than ground surface were used in the calculations. A more thorough discussion of stability factors appears in chapter five.

2.5.4 Wind Speed and Direction

The speed and direction of wind in alpine areas is known to be highly complex and instruments located other than at ridge-top locations rarely provide an accurate measure of the free movement of atmospheric air masses (La Chapelle 1970). Topographical obstructions and thermal effects (Geiger 1961) frequently distort the flow regime such that instrument readings are highly site specific. It was observed from the CF and SB wind data that large and irregular variations existed between the two recording sites, and hence precluded accurate extrapolation during periods of missing data. Wind direction at CF was also found to considerably vary from the flow of free air aloft, the major discrepancies being due to the valley location of the CF station. The most reliable record of free air direction was obtained during west to north-westerly airflow (the approximate aspect of the CF anemograph).

The 850 and 700 mb wind flow directions in radiosonde ascents from Christchurch were used to estimate general airmass directions. Any potential lag between Christchurch and the Craigieburn Range was considered whenever possible.

2.5.5 Precipitation Gauge Catch

Strong spatial variability of precipitation in rugged topography is well known [for example Grant and Rhea (1974); Hamilton (1962); Hovind (1965); Hendrick et al. (1979)]. Problems even exist on flat terrain in obtaining a representative gauge catch for any one location [Brown and Peck (1962); Goodison (1978); Rodda (1971)]. The reliability of gauge catch relates primarily to orifice dimensions and the degree to which airflow is properly directed over the gauge. The large orifice and weighing-bucket design of the Belfort gauges used in this study are thought to have minimized the effect of snow 'capping'. In cases where the SB gauge was bridged by snow, attempts were made to reconstruct the precipitation trace from the CF record.

2.5.6 Snow Pit Analysis

The instruments used in standard snow pit observations have not been significantly altered since the first large scale introduction of snow survey kits (Klein 1950). As the knowledge concerning snow on a micro-scale has increased, primarily because of laboratory research, the methods, instrumentation and classification techniques of snow observed in the field have lagged behind. For the purposes of this study two major improvements were made to the instrumentation usually employed in snow pit analysis. First, a PENTAX monocular (8 x 30 magnification) with a built-in 0.1 mm graduated scale was used for the identification of grain size and type. This author believes many inaccuracies have resulted in previous research which relied on a standard 1.0 mm grid and low powered magnifying lens for crystal classification. Second, a DIGITRON digital probe thermometer with a read-out scale defined in units of 0.1°C ($\pm$ 0.2% of 0.5°C output) was employed in the latter part of this study as the replacement to the less accurate bimetallic stem thermometers (approximate measurement accuracy of $\pm$ 0.25°C when accurately calibrated).

A discussion of accurate temperature and crystal measurement is included in chapter four.

2.6 REPRESENTATIVENESS - TIME AND SPACE VARIATIONS

Because of the extremes encountered in mountain environments and the site specific nature of many of the observations presented in this study, it was decided to include the following discussion concerning representativeness at a number of spatial and temporal scales.

2.6.1 Spatial Macro-Scale

Spatial variations at a macro-scale refer to differences on a climate or mountain range scale. Although the Craigieburn Range has been shown (section 2.3.3) to be representative of east leeward ranges of the central South Island, the extent to which SB is representative of the whole range must be considered. Observations by this researcher and personnel at some of the ski-fields along the Craigieburn Range suggests that spatial variations exist in the storm origin and amount of snowfall. In general, greater amounts are received in the northern end of the Craigieburn Mountains from north-westerly storms, including the study site at SB, while in the more southern reaches greater snowfalls originate from southerly directions. However, comparison of snow pit analysis along the eastern slopes of the range suggests that the snowpack structure at any given elevation is very similar, most of the variability existing because of differences in slope and aspect (discussed below). As yet there is insufficient quantitative data to support the validity of these observations but the avalanche prediction program planned for the area in 1982 will be gathering this type of information.

2.6.2 Spatial Meso-Scale

In terms of meso-scale representativeness two points will be considered, the location of the SB climate station and the snow pit analysis.

First, although the SB site is the second highest daily serviced climate station in New Zealand, it is still just within the major snow accumulation zones of the Craigieburn Range. An attempt was made to locate other instruments at higher alpine locations, specifically on Mt Cockayne, but as mentioned the harsh environment of the higher elevations frequently restricted necessary servicing. Hence, a very unreliable record of direct climate observations was obtained for areas primarily above 1550 m. Extrapolation of climate data was made into the higher elevations and these are considered reasonably accurate because of the extensive climate network, covering a range of lower elevations, on which the extrapolation was based.

Secondly, in regard to snow pit analysis, snowcover characteristics are known to vary considerably over short distances even on flat terrain and specifically between different surfaces. Gray (1979) reviews the main works dealing with snow accumulation and distribution for a variety of surfaces. In order to obtain accurate areal averages for snowcover characteristics, extensive sampling networks must be employed as suggested by Chemerenko (1973). However, rugged alpine terrain frequently limits the extent to which this can be achieved. Consequently, most studies in alpine areas have only considered areal averages of snow depth and density at a meso-scale and most other structural characteristics at a macro scale. Meso-scale variations in snowpack structure are known to be at an extreme in zones of contrasting aspects, primarily because of wind affected accumulation or differences in surface energy exchanges. Since this study focuses on the leeward areas of the Craigieburn Range, most extremes in snowpack structure are expected to be due to differences in energy exchanges, primarily controlled by radiation. The results to be presented from the snow pit analysis, which was conducted on contrasting sunny and shady slopes, may therefore be considered representative of extremes to be found at high elevations of the Craigieburn Range. The influences of spatial variations in snow depth or snowpack structure are further outlined in the main text.

2.6.3 Temporal Representativeness

As outlined in section 2.3.2, definitions of mountain climate in New Zealand are extremely tentative, primarily because of the scarcity and short term nature of alpine climate data. The description of the alpine snow environment in this study is only based on a five to six year climate record and two to three years of snow pit observations and therefore cannot necessarily be considered representative of long term trends or averages. In many cases, the results must be treated as temporally specific. The following attempts to evaluate the representative nature of the study periods with reference to past climatic information.

Over the very long term, as measured in thousands of years, a number of fluctuations have occurred in the New Zealand alpine climate. Burrows and Greenland (1979) have reviewed most of the available evidence concerning climatic trends since 1000 AD and point to numerous fluctuations in the snowline of the central South Island over the last millenium. The elevation of the annual snowline for the series of small glacial advances during this period was at a maximum of 200 m below the current limit of 1900 m. The present is considered to be an inter-glacial phase either due to reduced snowfalls and/or warmer temperatures.

The most reliable record of New Zealand climate dates back to approximately the mid-eighteenth century when regular climate observations were first begun. In an analysis of this record from lowland locations, Salinger and Gunn (1975) conclude that no significant temporal variations in precipitation have occurred although noticeable fluctuations are prevalent in temperature. A general warming has occurred since the 1850's with two intervening cold periods, one in the early 1860's and another from 1900 - 1935 [Salinger and Gunn (1975); Salinger (1976)]. Burrows (1976a) also attests to a colder climate in the recent past than at present, citing reports of icebergs near New Zealand in the 1850's and 1890's, and similarly Burrows (1976b) and Tomlinson (1970) have pointed to concentrations of severe snowfalls during two periods, 1860-1880 and 1920-1940. MacLeod (1974, 1980) in an historical account of the Grasmere Station, next to the Craigieburn Range, also notes major snowfalls in the past, specifically in the years

1918, 1939 and 1945. However, since the 1940's exceptional snowfalls have been rare and no icebergs have been seen in southern oceans near New Zealand.

The years since 1950 have been reported by Salinger (1979) to be the warmest on instrumental record although most of the variations responsible for the warming have occurred in the summer rather than winter months. During the last forty years numerous glaciers of the South Island have shown massive retreats [Burrows (1977); Salinger (1976); Wardle (1973)] and more particular to this study, Canterbury glaciers have experienced accelerated retreat since 1950 [Burrows (1973); Burrows and Maunder (1975)]. Although no direct scientific data have been presented to the fact, Burrows and Greenland (1979) believe that over the last seventy to eighty years the snow line of the Craigieburn Mountains has receded.

Morris and O'Loughlin (1965) and O'Loughlin (1969a) have defined the more recent temporal trends in the snow environment of the Craigieburn Range. Figure 2.11 presents a recent update of snow surveys which were conducted near the study site on which this report is based. Generally, from 1962 to 1973 large fluctuations existed in the amount of water accumulated in the snowpack both on a monthly and annual basis. In particular, 1968 was noted as being a heavy snow year, both in terms of total accumulation and duration of snowcover.

There exists no overlapping snow accumulation records between the periods 1962-1973 and 1975-80 but it was hoped to make a comparison of the two indirectly through the use

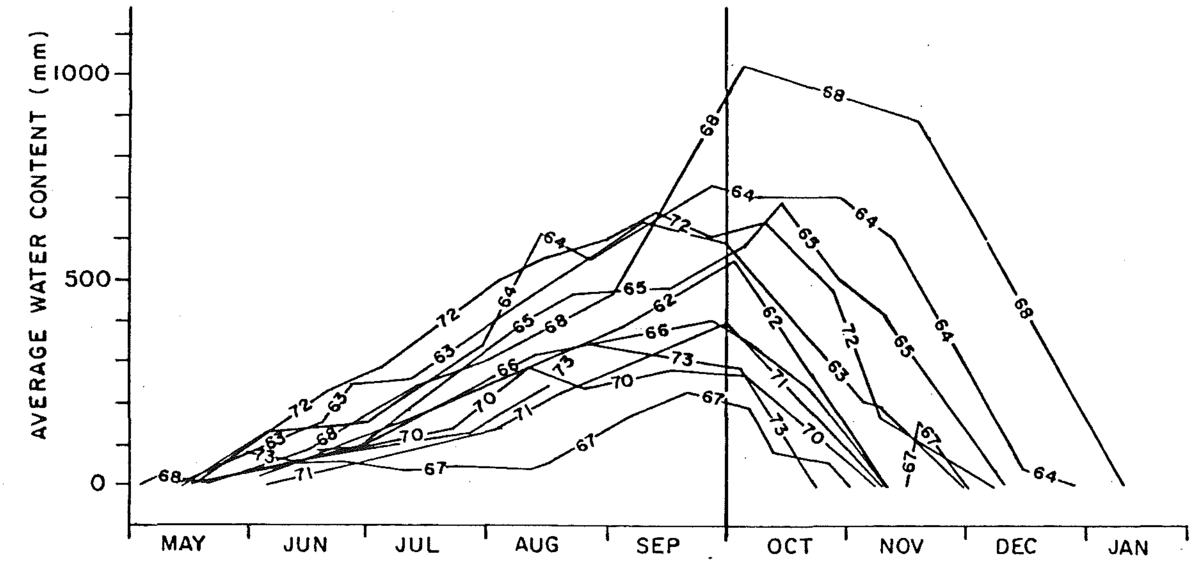


Figure 2.11 SNOW SURVEY RESULTS ALLAN'S BASIN (AFTER O'LOUGHLIN).

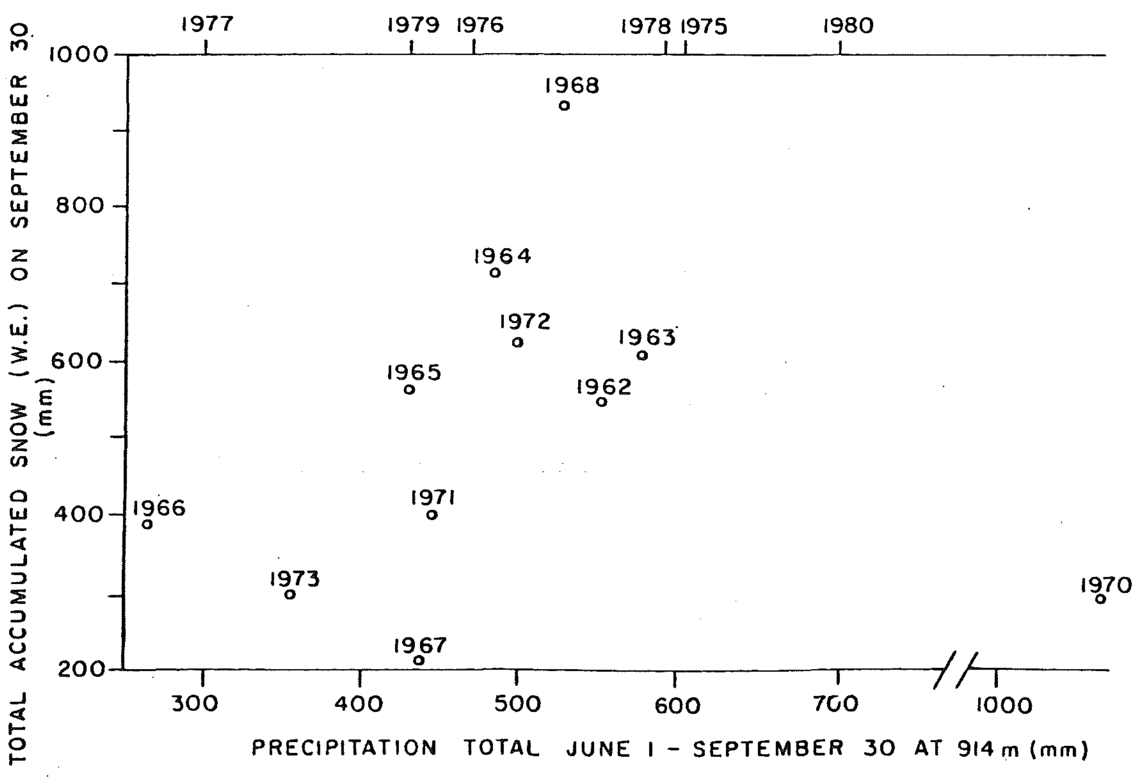


Figure 2.12 WINTER PRECIPITATION TOTALS - SNOWPACK WATER EQUIVALENT.

of winter precipitation totals. Figure 2.12 was therefore constructed, and compares the end of September snowpack water equivalents for 1962-1973 with the total winter precipitation (June-September) measured at the CF station for the same years plus those from this study. Unfortunately, there is a very poor correlation between the snowpack water equivalent and CF precipitation figures. Thus, high winter precipitation is not necessarily reflected in large snow accumulation. The years 1962, 1963 and 1970 recorded greater winter precipitation than 1968 but all had much lower total snow accumulations. The variables most likely responsible for the poor correlation are temperature, through its effect on winter melt and precipitation type (rain-sleet-snow), and wind redistribution. All these factors are discussed in later chapters.

The poor correlation precludes a comparison of total winter snowfall between the 1962-1973 and 1975-1980 years, but as indicated in figure 2.13, the total range in winter precipitation is similar between the two sets of years.

In a further attempt at assessing the temporal representativeness of the 1975-1980 years, mean monthly values of total precipitation, wind speed, temperature and daily radiation were compared to their respective long term means as described in section 2.3.3 (appendix B). Generally, precipitation demonstrated the most marked variations, fluctuating $\pm$ 100 mm per month about the long term mean. Variations in the other parameters were approximately $\pm$ 1 m s^{-1} for wind speed, $\pm$ 2 - 3^{o}C for air temperature and $\pm$ 2 MJ d^{-1} for global radiation.

In summary, it would appear that the mountain climate of the Craigieburn Range during the period of this study, 1975-1980, is reasonably representative of conditions prevailing over the last two decades, although no year could be said to have experienced exceptional snowfall. Over the longer term, the period from 1950 may be considered to be one of light snowfall and slightly warmer temperatures, especially in comparison to the latter part of the nineteenth century and the beginning of this century.

In terms of temporal representativeness, one other point should be considered, the frequency of sampling. A majority of the snowfall and snowmelt results are based on continuous monitoring of the relevant climatic variables. However, in the case of snowpack structure, sampling was only conducted on a weekly to bi-weekly basis. As will be described in chapter four, appreciable changes occurred in snow structure within time frames of less than a week, in some cases during only a few hours. Continuous monitoring of snowpack structure was not practical nor feasible during some of the storm cycles which produced the most rapid structural changes. Instead, supplementary meteorological data were used in conjunction with the weekly snow pit observations to detail what were considered the major temporal variations in snowpack structure.

CHAPTER III

SNOWFALL CHARACTERISTICS

3.1 INTRODUCTION

Information concerning the characteristics of snowfall in New Zealand is extremely limited. Previous meteorological studies of snow storms have primarily focused on infrequent events which were classified as 'exceptional' because of their impact on lowland and sub-alpine areas where large snowfalls do not regularly occur [Hughes (1969, 1974); Neale and Thompson (1977); Tomlinson (1970)]. Within the alpine zone, Morris and O'Loughlin (1965) and O'Loughlin (1969a) are the only ones to have considered any long term trends in snowfall, while Fitzharris (1976), McNulty and Fitzharris (1980) and Weir and Owens (1981) have detailed individual storms. A simple synoptic classification of snow storm types has also been devised by Owens and Prowse (1980).

However, there still remains a lack of information dealing directly with long term snow storm and inter-storm snowfall characteristics. This chapter describes a number of snowfall characteristics for the Craigieburn Range.

Snowfalls are regularly defined in specific manners for a number of practical purposes such as accumulation studies, avalanche prediction and snowmelt hydrology. The sections in this chapter detail specific characteristics of snowfall at the SB station in view of these practical considerations. A total of seven results sections are presented following a review of the data base and methodology used in the analysis. The first section describes the composition of yearly and monthly precipitation. The second considers snow storm events and outlines information concerning storm magnitude, duration and intensity. The role of elevation in controlling the above parameters is analyzed in the third section of results, which is followed by a discussion of snow storm temperatures and freezing levels. The general synoptics of large snowfalls is then discussed and an analysis of new snow density values concludes the results sections. A brief summary at the end of the chapter reviews the major findings from each section.

3.2 DATA BASE - METHODOLOGY

The precipitation data used in this study were primarily recorded at the SB meteorological station, but other data from the CF station were also employed for the purposes of confirmation and in the section concerning the influence of elevation. All data interpretation from the chart form was performed manually to ensure maximum accuracy and validation. Since the completion of this report many of the chart data have since been digitized and now exist on computer file. However, this researcher believes serious misinterpretation of the results can occur if the digitized data are used to the complete exclusion of the original chart recordings.

Separation of the precipitation record into categories of snow, sleet and rain was performed over a five year period from March 1975 to February 1980. A reliable precipitation record was considered to exist from March 1975 onwards. The date of February 1980 was used as an ending date of the analysis to provide 5 full yearly records. Within the text of this chapter a referenced year will refer to the period from March of that year to February of the following year. Ideally, continuous visual observations of precipitation forms is the most reliable method of classifying all precipitation but practically this type of approach is not feasible. The daily climate observations at both the CF and SB stations contained descriptions of precipitation forms, periodic storm snow line elevations and records of both new and total snow depth. These were used in the classification of precipitation but since so many storms were known to contain a variety of precipitation types, a supplementary classification technique, based on air temperatures at the time of precipitation, also had to be employed.

Lowndes et al. (1974) has found there is an eighty percent probability that precipitation will arrive as snow if the freezing level is at the ground surface. Frequent observations during the 1978-80 winters in the Craigieburn Range revealed that precipitation was exclusively in the form of snow when the air temperature was below freezing, and at temperatures above 2°C, any solid form of precipitation was relatively uncommon. Manley (1970) has noted that the wet bulb air temperature at the surface rarely exceeds 3°C during periods of snow or even sleet. There do exist some reports of snow and sleet occurring in quite warm air temperatures but these are usually rare and short-lived phenomena. The brevity of such events is due to the cooling effect of the falling snow on the air through which it passes (Lumb 1961). Any appreciable amount of falling snow will rapidly lower the air temperature to, or below, the freezing point.

Based on the above, it was decided to separate the precipitation record into three categories according to the concurrently recorded air temperature record. Whenever possible the daily climate observations were also employed to verify the classification. Below a 0°C air temperature, all precipitation was regarded as snow and above 2°C as rain. It was believed the greatest irregularity in precipitation forms occurred between 0 and 2^{o}C but unless some form of visual observation had been made the precipitation occurring in this range was regarded as sleet.

In the use of near surface air temperatures for classifying precipitation, care had to be taken to ensure the temperatures were representative of the general free air. Thermal inversions may develop over a snow surface, especially during periods of minimal surface turbulence, and may result in a generally lower temperature being recorded at the standard instrument height than that for the free air. In periods when this problem was thought to exist, verification of the screen air temperature at the SB site was made using environmental lapse rates derived from thermograph stations at lower elevations in snow free conditions.

The actual identification of some precipitation events was also complicated by the presence of blowing snow during precipitation-free periods. Comparison of the SB precipitation record to that of the relatively wind sheltered and snow free CF station, plus checks of the daily climate observations were believed to eliminate this problem.

Information concerning precipitation intensity was also directly derived from chart records. At any visible change in the slope of the precipitation trace, new intensity values were calculated. Over the five year period approximately 1900 separate intensity readings were made. Problems were known to occur when the gauge became capped with snow but attempts were made to reconstruct such events from the CF data whenever possible. The full extent of this problem is unknown but the effect on the total precipitation record is considered to be small.

Individual storm events were originally separated from the total precipitation record whenever any precipitation was followed by at least twenty-four hours of dry conditions. This time interval was the same as that used by O'Loughlin (1969a) in his study of Craigieburn storms. However, after a consideration of the synoptic storm patterns over the 1975-80 period, a twenty-four hour separation was found to be too long, resulting in the grouping of distinctly different storm systems. A second separation was then completed based on an eighteen hour period. This time frame seemed to coincide much more satisfactorily with the synoptic storm patterns.

3.3 YEARLY AND MONTHLY PRECIPITATION RESULTS

The first row of figures in table 3.1 lists the total yearly precipitation figures for the 1975-80 study period. The overall mean total of 1433 mm (maximum error of estimate = $\pm$485 mm at 95% confidence level) is only 153 mm different from the long term mean (1966-1979) reported by McCracken (1980a). The extreme yearly variations from the mean (-39% for 1977 and +36%) for 1978 are also well within the average variation reported for the area by Rowe (1970), and Seelye (1940) for the inland strip of Canterbury. Hence, the 1975-80 totals can be considered to be reasonably representative of approximately the last two decades.

PRECIPITATION (mm)	1975-76	1976-77	1977-78	1978-79	1979-80	MEAN	MAXIMUM ERROR OF ESTIMATE AT 95% CONFIDENCE LEVEL
TOTAL	1566.3	1310.8	873.7	1467.3	1945.0	1432.6	±484.2
SNOW	407.9	334.3	272.0	430.4	437.7	376.5	±88.5
SLEET	279.4	176.4	183.5	220.6	189.4	209.9	±52.4
SNOW + SLEET	687.3	510.7	455.5	651.0	627.1	586.3	±122.4
RAIN	879.0	800.1	418.3	816.2	1317.9	846.3	±397.2

Table 3.1 ANNUAL PRECIPITATION TOTALS. Totals are accumulated from March in the first year to February in the second.

The annual totals of the three classes of precipitation, snow, sleet and rain, are listed in the bottom four rows of table 3.1. On average snow contributed at the SB site 377 mm yr^{-1} or 26% of the total precipitation over the five year period. The overall range in the percentage of total precipitation as snow was small, varying from only 23% (438 mm) in 1979 to 31% (272 mm) in 1977. These are the same years which recorded the greatest and least total precipitation respectively. If the sleet category of precipitation is added to the snowfall totals, the mean percentage of total yearly precipitation rises to 41 with 32 and 52% being the extreme values.

The mean percentage values of 26 for snow and 41 for snow plus sleet are only marginally different than the 33%

estimate made by Morris and O'Loughlin (1965) for snowfall as a percentage of annual precipitation above 1500 m. This similarity in results would tend to indicate that the temperature based separation of precipitation is a reasonably accurate technique.

Based on the assumption that the actual annual snowfall at SB is between the mean calculated values of 377 m for snow and 586 mm for snow plus sleet, comparisons may be made to snowfall totals in other climates. The SB snowfall inputs are considerably less than those expected for regions, especially at higher elevations, in closer proximity to the west coast of New Zealand. Similarly, they cannot be considered representative of west coast alpine conditions of North America as defined by La Chapelle (1966) and Fitzharris (1975). In direct contrast, however, these amounts are three to four times the total annual precipitation for the interior Arctic climate (Thompson 1967) and approximately equal to the total yearly precipitation in much of the continental regions of North America and Central Asia [Bradley (1976); Lockwood (1974)]. It would appear that the SB snowfall inputs are more typical of the intermontane regions of North America as described by La Chapelle (1966).

The actual distribution of snow throughout the year at 1500 m is quite varied as evident in table 3.2 and appendix C. As observed by O'Loughlin (1969a) all forms of precipitation may occur in any month of the year. August, however, is the month which was found in this study to record the highest mean percentage of precipitation

		JAN	FEB	MAR	APR	MAY	JUNE	JULY	AUG	SEPT	OCT	NOV	DEC
SNOW	MEAN	0.2	12.2	0.6	6.6	18.0	36.5	58.3	88.4	38.1	33.9	12.4	4.7
	RANGE	0.0-0.7	0.0-36.4	0.0-2.5	0.0-18.1	7.3-56.3	2.9-49.0	46.4-76.5	65.5-93.8	19.0-79.0	24.2-50.0	41.0-36.4	0.0-21.2
SLEET	MEAN	4.2	9.5	11.6	10.3	9.3	30.6	25.7	8.5	19.9	17.9	15.4	14.7
	RANGE	0.8-7.8	0-32.6	3.4-64.4	6.0-15.6	1.4-26.9	6.3-59.0	13.4-44.0	0.8-33.8	17.7-29.7	7.9-43.1	3.9-24.5	7.6-28.6
SNOW + SLEET	MEAN	4.4	21.8	12.2	17.0	27.3	67.0	84.0	96.9	58.0	51.8	27.9	19.4
	RANGE	0.8-8.6	0.0-36.8	3.4-64.4	7.4-24.0	13.0-59.6	28.4-88.0	59.8-100.0	94.9-99.4	40.4-98.4	38.3-70.4	16.1-59.7	12.4-36.5
RAIN	MEAN	95.6	78.2	87.8	83.0	72.7	33.0	16.0	3.1	43.0	48.2	72.1	80.6
	RANGE	90.6-99.2	31.1-100.0	35.6-96.6	77.6-92.6	40.4-87.0	12.0-71.6	0.0-40.2	0.6-5.4	1.6-63.3	29.6-61.9	40.3-83.9	71.4-87.6

Table 3.2 MEAN PERCENTAGES OF MONTHLY PRECIPITATION AS SNOW, SLEET AND RAIN. Mean percentages are derived from the five year precipitation record. The range values refer to the maximum and minimum percentages over the 1975-1980 period.

occurring as snow (mean = 88.4%). On a yearly basis snowfalls are also greatest in total quantity during the month of August, with an average input of 114 mm and a maximum of 185 mm.

The presence of rain during the winter months is, as will be described in chapters four and six, a significant factor in the structural development and ripening of the Craigieburn snowcover. Although all months over the five year period recorded some rainfall, it was August which had the lowest monthly average of only 3.1%, varying in extremes from 0.3 to 5.4%, and never exceeding a monthly total of 11.0 mm. Considerably greater percentages of monthly precipitation are contributed in the form of rain for the months on either side of August, the mean percentages being 16 and 43 for July and September respectively.

The presence of rainfall during the main winter months is a regular feature in maritime zones of seasonal snowcover. Benson (1979); Fitzharris (1975) and La Chapelle (1966), working in different regions of the maritime climate of North America, all report the regular winter time presence of rainfall. In particular, Fitzharris (1975), found that snow contributed only 76% of the winter precipitation (December to March) at an elevation of 1260 m on Mt Seymour in British Columbia. The amount and frequency of rainfall during winter has also been observed to decrease with distance from coastal areas such that both Benson (1979) and La Chapelle (1966) have introduced a coastal-transition zone to separate between the coast-maritime and cold continental snow regions. As yet no quantitative definition, in terms of precipitation magnitude or composition, has been introduced for such a classification. The above results suggest that a coastal-transition climatic description might be appropriate for the Craigieburn Range but this point is discussed in more detail in later sections.

It is noteworthy that although strong inter-year variations are apparent in total and monthly precipitation figures, much of the variability is attributable to single relatively short term events. To exemplify this point, 75% of the 247 mm of snow which fell during the June to August period in 1979, occurred in one three day event at the end of August. The following section deals with the comparative size and duration of individual snow storms.

3.4 SNOW STORM MAGNITUDE, DURATION AND INTENSITY

The magnitude and intensity of snowfalls in alpine areas are critical parameters to the eventual metamorphism and stabilization of snowpacks. Snowcover produced from large but infrequent snowfalls is known to have a much more homogeneous stratigraphy than one which has developed from smaller but more frequent snowfalls. In the latter case, many more layers are exposed for extended periods to surface energy exchanges which produce more extensive and varied structural changes in the ice skeleton than those which normally occur in the sub-surface layers of a snowpack.

The snow storms used in the following analyses were selected from the results of the eighteen hour dry period separation. Only events which produced a minimum of 5 mm of precipitation in the form of snow, and for which snowfall was the dominant precipitation type, were included. A total of eighty-four events were identified, the frequency of which is illustrated in figure 3.1. The mean storm duration was calculated at approximately 46 hours. This value is appreciably larger than that identified by O'Loughlin (1969a); the discrepancy in results probably due to the shorter time frame used to separate events. However, the modal class (based on a ten hour class size) lying between 21 and 30 hours is considerably less than the mean value of 46 hours. It would appear from these results that most snowfalls occur over reasonably lengthy periods with many events lasting two or more days.

Since most events seem to be of a generally lengthy nature, the critical factor for total accumulation is how much precipitation occurred during each storm interval. Figure 3.2 displays the total precipitation recorded during each of the eighty-four events. The mean event magnitude is 21.8 mm although there is a definite skew to the distribution. Approximately fifty percent of the events recorded a total precipitation of less than 15 mm and over one third had less than 10 mm. Fourteen events with total amounts greater than 40 mm were largely responsible for the high mean value.

Storm magnitude has been suggested in the United States by Judson (1967) to be a useful measure in the prediction of snow slope failure in alpine areas. A storm precipitation total of 25 mm was deemed critical for avalanche generation. In reference to figure 3.2 twenty-one

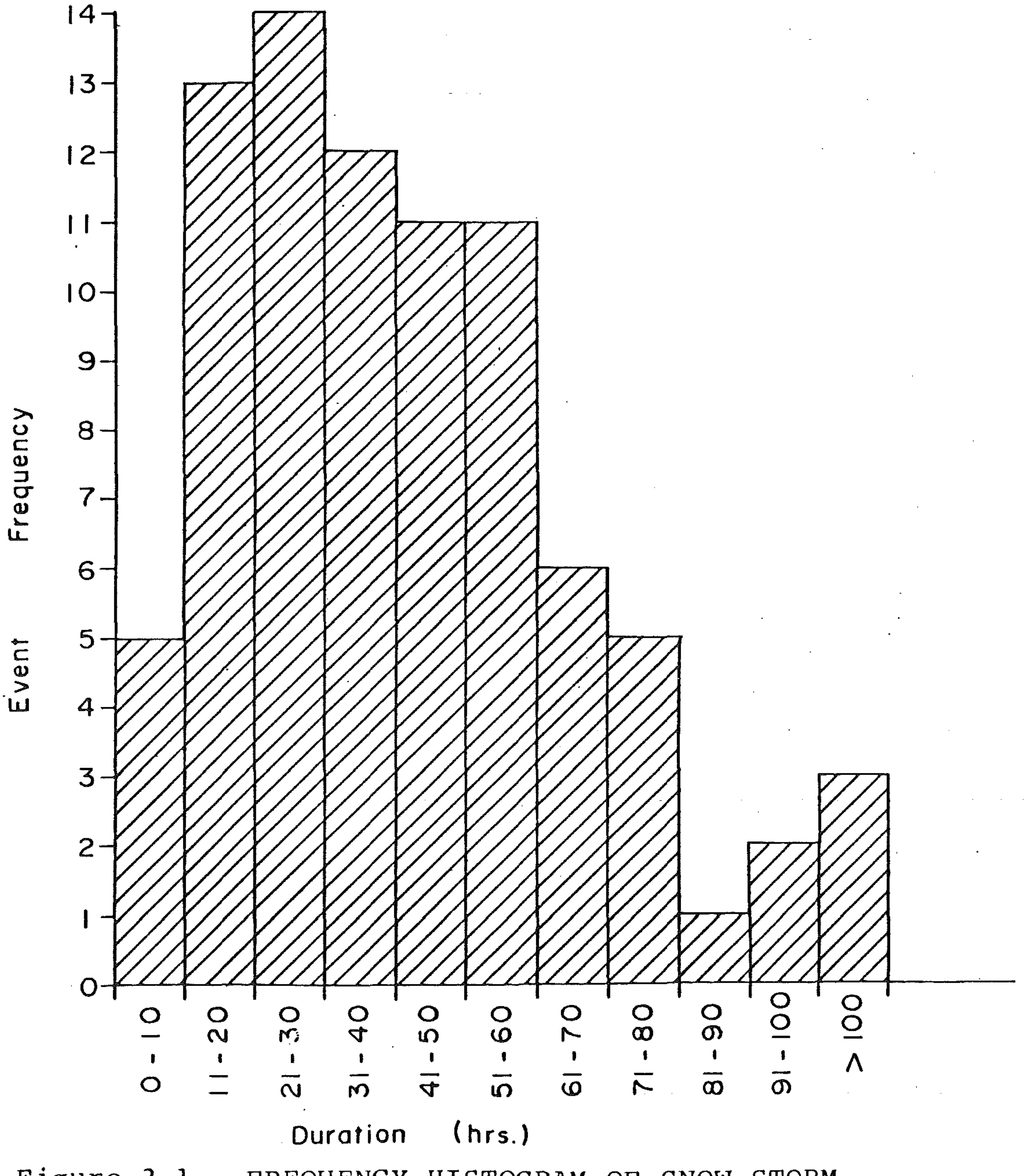


Figure 3.1 FREQUENCY HISTOGRAM OF SNOW STORM DURATION.

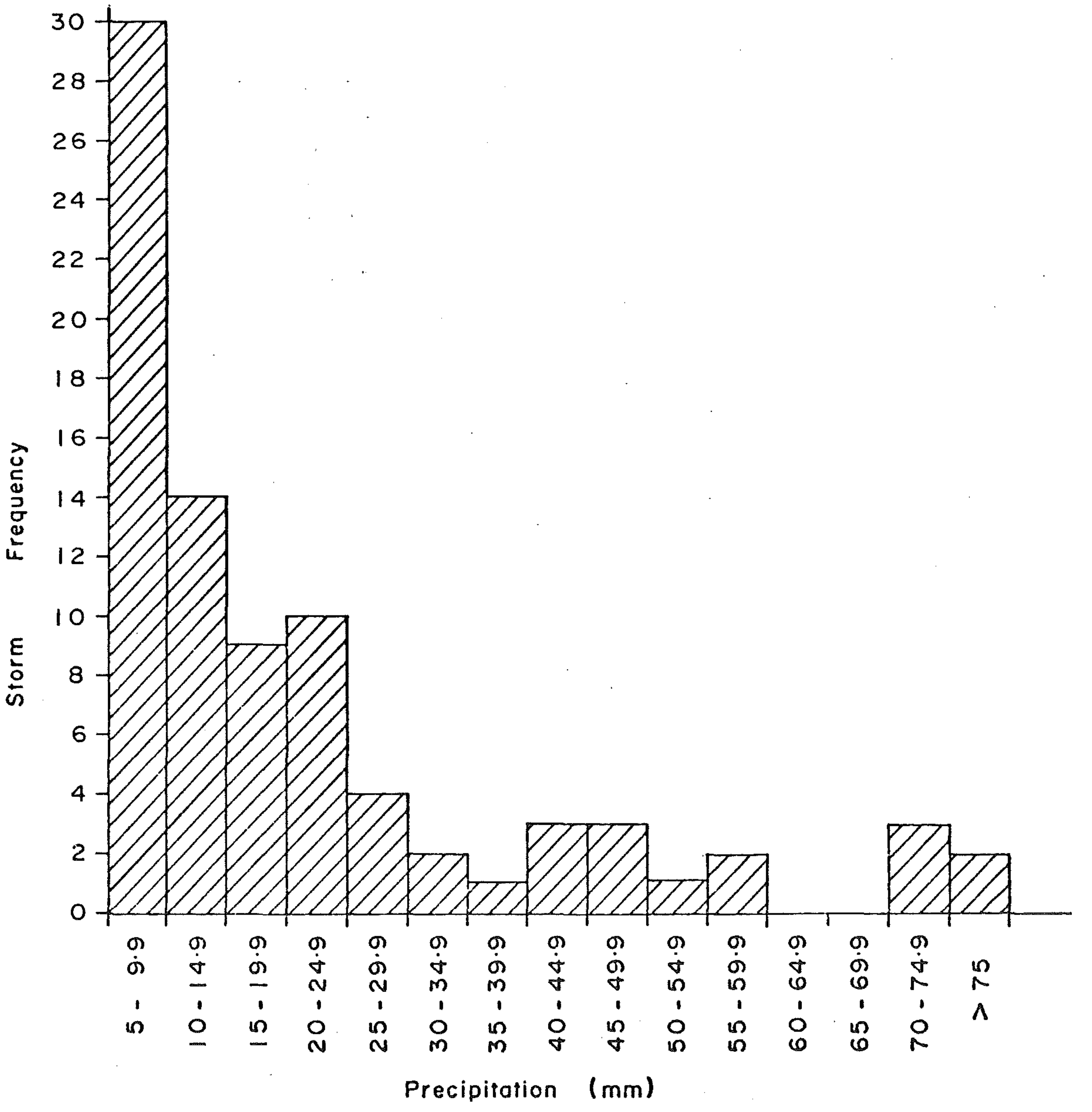


Figure 3.2 FREQUENCY HISTOGRAM OF SNOW STORM PRECIPITATION TOTALS.

of the storms produced 25 mm or more of total precipitation but only 15 recorded this amount while the air temperature remained below the freezing point. However, as illustrated in figure 3.3 all but one of any of these storms required forty hours or more to reach the critical accumulation value of 25 mm. Such lengthy periods are likely to allow considerable settlement and stabilization (processes described in chapter four) of new snow, especially if air temperatures are near the freezing mark. Hence, over entire storm periods the total accumulated snowfall is unlikely to pose any serious avalanche risk.

The above discussion of the duration-magnitude relationship assumed that precipitation was evenly distributed throughout the individual events. In most cases, precipitation intensities were found to vary considerably during storm events and as noted by Morris (1965), the highest intensities are from north-westerly storms.

To evaluate long term patterns in snowfall intensity, a cumulative intensity diagram (figure 3.4) was constructed based on all precipitation from the eighty-four events which occurred at sub-freezing air temperatures. Atwater (1952) has suggested that a snowfall rate of 2.5 mm hr^{-1}, especially in the presence of a high wind, should be considered as a critical loading rate in snow slope failure. As illustrated in figure 3.4 26% of the total snowfall occurred at an intensity greater than 2.5 mm hr^{-1}. In comparison, the individual mean and mean maximum storm event intensities were 0.8 and 2.8 mm hr^{-1} respectively. However, a high

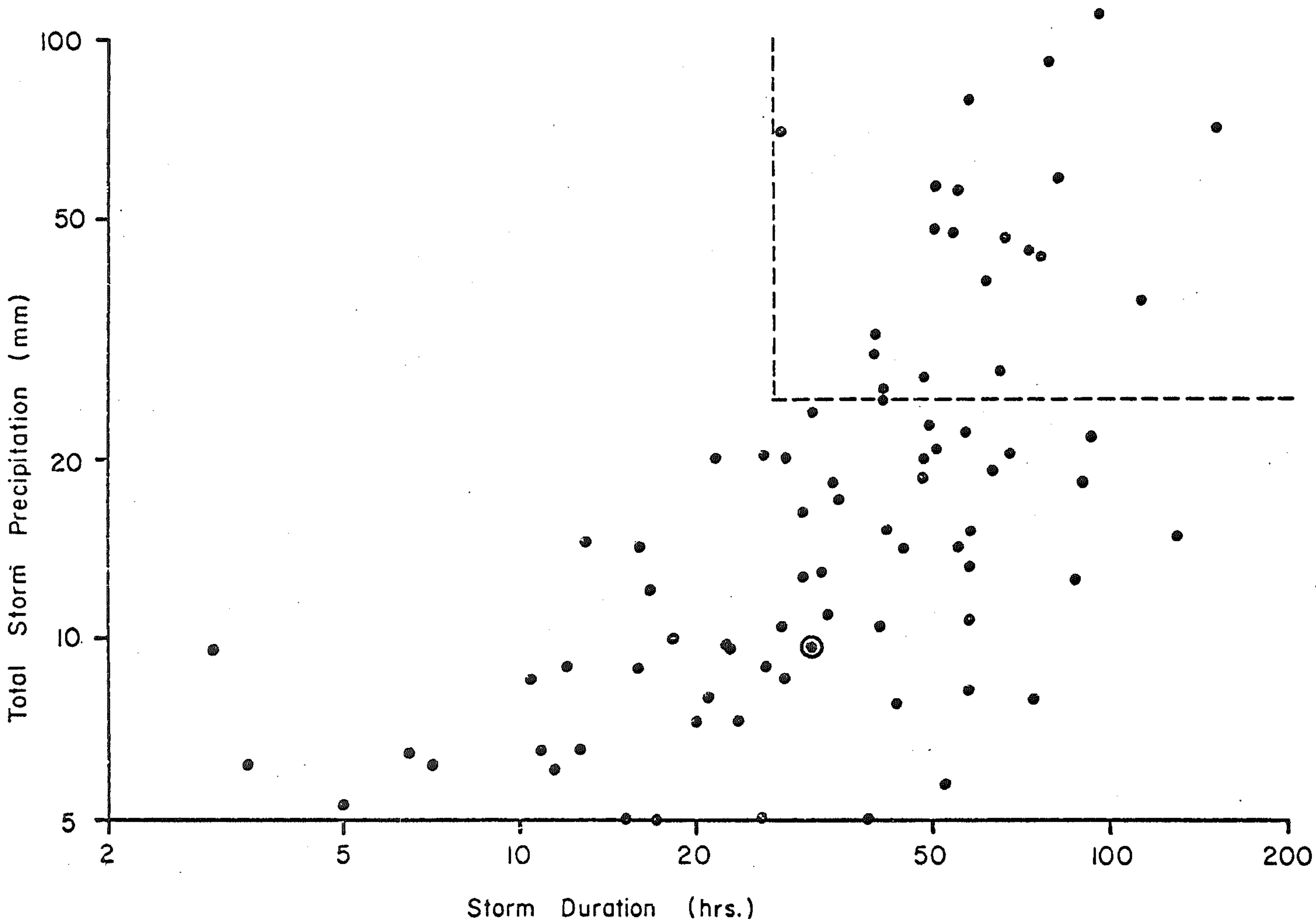


Figure 3.3 RELATIONSHIP OF STORM DURATION TO STORM PRECIPITATION TOTALS. Boxed area indicates storms with more than 25 mm of precipitation.

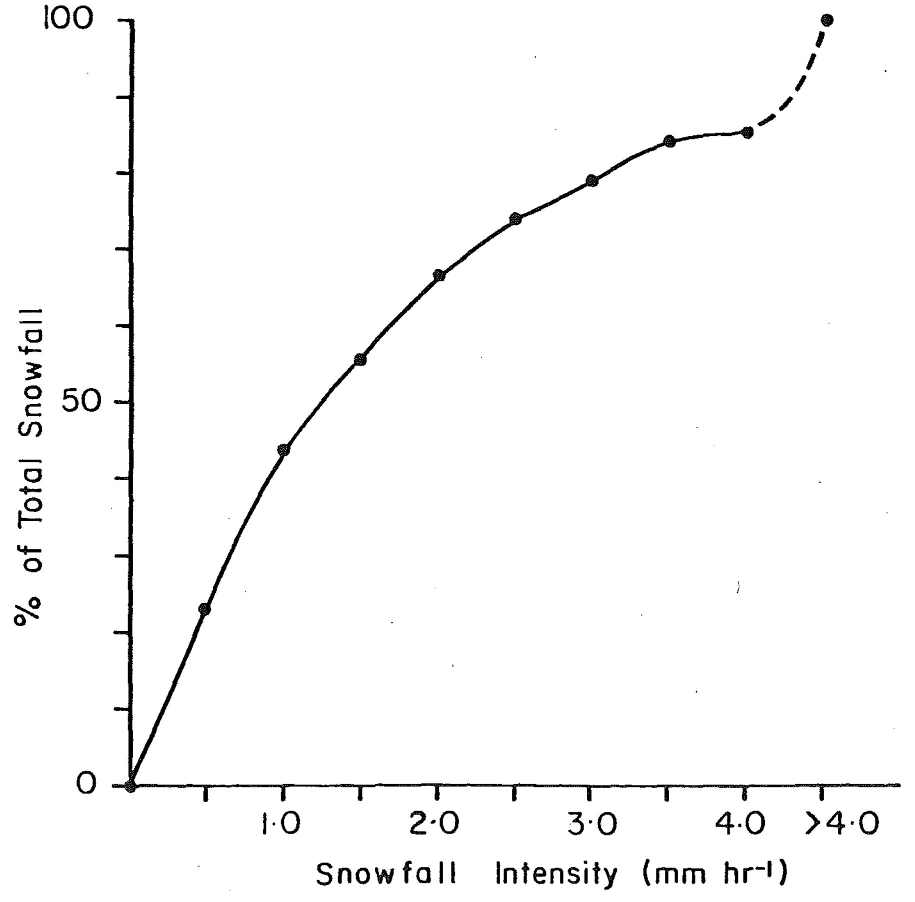


Figure 3.4 CUMULATIVE SNOWFALL INTENSITY CURVE.

intensity snowfall can only be considered dangerous in terms of snow slope failure, if it persists long enough to produce appreciable accumulations of snow.

By combining the results from the intensity and accumulation-event analysis, five storm events were revealed to produce greater than 25 mm of snow, and ten storms greater than 10 mm, at a rate exceeding 2.5 mm hr^{-1}.

In summary, most snowfall events are characterized by low intensities although there is an average of one storm per year, which in terms of avalanching, produces enough snow at a sufficiently high rate that safe settlement and stabilization of snow slopes is unlikely. Such storm conditions cannot be considered typical of maritime snowpacks where, as noted by La Chapelle (1966) in the United States, large quantities of snow fall in single storms and snowfall intensities may reach 300 mm hr^{-1} (snow). For the Craigieburn region, O'Loughlin (1969a) has estimated that the mean snowfall intensity during storms is only 23 mm hr^{-1} (snow). Similarly, based on the results from the precipitation-intensity analysis and an average new snow density of 150 kg m^{-3} (described in a later section) mean storm event intensities would only be 12 mm hr^{-1} (snow).

3.5 EFFECT OF ELEVATION ON SNOWFALL, MAGNITUDE, DURATION AND INTENSITY.

The previous sections have focused on a precipitation record from one elevation in the Craigieburn Range. For most mountainous regions precipitation is known to increase with elevation and hence many precipitation characteristics from a single elevation do not universally apply for all elevations. Information concerning elevational relationships is extremely valuable, especially in avalanche forecasting where the magnitude and intensity of high elevation snowfall must be predicted from data collected at low level stations.

A number of snow accumulation studies in New Zealand by Fitzharris (1972), Chinn (1969) and Weir (1967) have pointed to the roles of both precipitation and melt in the development of characteristic snow wedges over a range of elevations. As earlier mentioned, McSaveney (1978) also points to the precipitation profile across New Zealand. However, no published studies are available which directly consider relationships between elevation and precipitation in the form of snow. This section focuses directly on those relationships.

As identified by Morris and O'Loughlin (1965) over eighty percent of the precipitation at the SB site originates from the south-west to north-west quarter. The Craigieburn Range is also known to be a considerable distance eastward of the point of maximum precipitation along the Southern Alps (McSaveney 1978). Hence, the eastern portions of the Craigieburn Range, including the study site, must be considered to be primarily in a downwind-rainshadow location. This type of location may not produce the same elevation-precipitation characteristics as would be expected on the west coast of New Zealand where orographic effects are most pronounced.

In the case of orographically produced uplifting the amount of precipitation at any elevation is dependent on the total amount of condensation from water vapour which is in turn proportional to time, pressure and more importantly vertical velocity and specific humidity. Within the range of elevations of most New Zealand mountains normal decreases in specific humidity with elevation should be compensated for by the increase in wind velocity. As propounded by Collier (1975) and Bader and Roach (1977), knowledge of the wind and moisture structures over any topographic barrier will allow the accurate modelling of orographic precipitation. This type of information is rarely available and is difficult to collect, hence most relationships between precipitation and elevation must be derived from direct observations of the precipitation distribution.

In an attempt to identify some of the relationships between snowfall and elevation for the Craigieburn Mountains, thirty-three snow events were selected from the 1975-80 precipitation records at the SB and lower elevation CF stations. Many of the original eighty-four storms were not included because of minor interruptions in the precipitation records, suspected gauge catch inaccuracies (a more rigorous definition was employed than in the larger eighty-four event sample because of minor differences between the two elevations), and where the beginning and end points for storms at both locations could not be accurately defined. In some cases, this latter problem eliminated storms with very gradual changes in precipitation intensity at either end of the event.

Although the sample set used in this study is much smaller than the original storm sample, it is still believed to be representative of most snowfall conditions. The mean amount of precipitation per event was found to be only 2.3 mm less than the overall mean of 21.8 mm mentioned in section 3.4.

The total amount of precipitation recorded at the CF and SB stations for the thirty-three storm events are pictured in figure 3.5. In general, the total amount of precipitation is greater at the higher elevation site for the full range of event magnitudes. A difference of means test (correlated samples at 95% confidence level) also revealed a significant difference between the mean storm magnitudes at each site. The results of the test appear in row one of table 3.3.

By regressing the SB storm precipitation amounts on the CF values the following equation was derived:

$$\mathrm{LOG}\ (\bar{P}_{SB}) = 0.85\ \ \mathrm{LOG}\ (\bar{P}_{CF}) + 0.27 \qquad (3.1)$$

where $\bar{P}$ is the total storm precipitation (mm); SB, CF refer to the SB and CF meteorological stations.

The data were log transformed to satisfy the assumptions of the regression analysis. As might be expected because of collinearity between the variables a relatively high correlation coefficient of 0.90 resulted.

A distinctive feature of figure 3.5 is the range in the ratios of storm precipitation at the two sites. In many cases, an approximate one to one ratio is present. Williams and Peck (1962) have suggested a small ratio may be

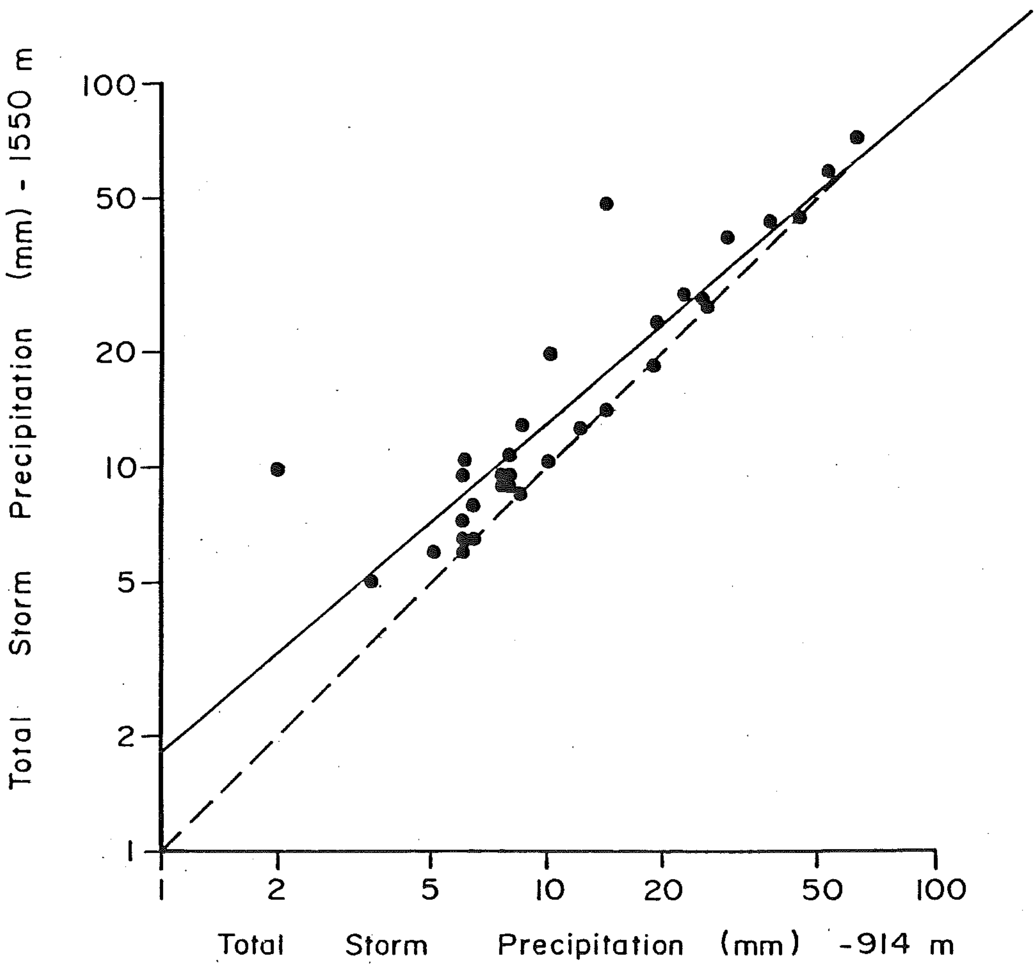


Figure 3.5 RELATIONSHIP OF STORM PRECIPITATION TOTALS BETWEEN THE SB AND CF CLIMATE STATIONS.

MEAN STORM EVENT CHARACTERISTICS	SB 1555 m	CF 914 m	t VALUES DIFFERENCE OF MEANS (95%)
TOTAL PRECIPITATION (mm)	19.42	15.50	4.32
EVENT DURATION (hr)	38.86	31.62	11.35
PRECIPITATION DURATION (hr)	29.10	20.29	4.59
EVENT INTENSITY (mm hr^{-1})	0.53	0.52	0.25
PRECIPITATION INTENSITY (mm hr^{-1})	0.71	0.77	1.24

Table 3.3 COMPARISON OF STORM CHARACTERISTICS BETWEEN THE SB AND CF CLIMATE STATIONS. Underlined values in column four refer to a significant difference existing between the means in columns two and three (95% confidence level).

generated when precipitation has little dependence on orographic or frontal lifting. In such cases, precipitation occurs from large scale upward motion associated with upper air 'cold lows'. The synoptics of some of the major snowfalls are discussed in section 3.8.

If the precipitation at the SB and CF is primarily of an orographic origin then, as outlined by Hendrick et al. (1979), increases in total storm precipitation with elevation should be the result of increased precipitation intensity. In order to test this relationship, the storm intensities were calculated for both the SB and CF stations. As some of the storms contained precipitation-free periods, intensities pertaining only to precipitation periods within the events were also calculated. Scattergrams for the two types of intensities appear in appendix D.

Regression lines have been calculated for the data but, as might be expected from the broad scatter of points, coefficients of determination of only 0.36 and 0.37 were derived. A comparison of mean intensities between the two sites appears in table 3.3 and a summary of the regression equations is listed in the first two rows of table 3.4.

In 40% of the cases examined, higher mean storm intensities were observed at the lower elevation station. This percentage increases to 65% if precipitation-free periods are eliminated from all events. Greater precipitation intensities at the lower elevation station have also been observed by Morris and O'Loughlin (1965). Although this inverse relationship may be related to inaccuracies of gauge catch it may also be explained by the presence of low cloud bases in the area. In the absence of updrafts, the cloud base is the elevation at which maximum precipitation for a given period should occur. Hence, when the cloud base is near the valley elevations, greater precipitation intensity would be expected at the CF rather than the SB station. This author has observed on occasion cloud cover and precipitation confined to elevations below 1500 m, while above this elevation clear skies dominated.

Since a large number of storms recorded greater intensities at lower elevations, there must be another factor

REGRESSION ANALYSIS BETWEEN THE SB AND CF STATIONS FOR:	REGRESSION AND CORRELATION COEFFICIENTS		
	a	b	r
EVENT INTENSITY	0.24	0.53	0.60
PRECIPITATION INTENSITY DURING EVENTS	0.21	0.64	0.61
EVENT DURATION	6.47	1.03	0.95
PRECIPITATION DURATION DURING EVENTS	0.35	1.42	0.94

Table 3.4 REGRESSION ANALYSIS OF STORM INTENSITY AND DURATION BETWEEN THE SB AND CF CLIMATE STATIONS.
Results are based on the regression equation Y = a = bx. The SB station was employed as the dependent variable. The original data with the respective regression lines appear in appendix

for the greater storm precipitation at higher elevations. The only other explanation lies with the actual storm duration.

For the thirty-three storm samples, storm duration and storm precipitation duration (excludes inter-storm precipitation-free periods) statistics were compared. Plots of the data appear in appendix D, and a summary of the regression equations in table 3.4. In all but a very few cases storm duration and storm precipitation duration were both greater at the SB station. The mean values for both duration measurements from each station were also compared and proved to be significantly different (table 3.3).

Figure 3.6 illustrates the differences between the mean values for storm duration and storm precipitation duration at the two elevations. Notably, the mean storm duration at 914 m is even less than the mean storm precipitation duration at 1555 m.

Hendrick et al. (1979) working in the north-eastern United States have also reported that on an event basis, it was duration which explained a seasonal increase in precipitation with elevation. They concluded that similar intensities at both upper and lower sites was attributable to the presence of low intensity precipitation, mainly at either end of the main storm event, which was not concurrent with low level precipitation. Hence, precipitation intensities averaged over the individual storm periods would be similar to those at the lower elevation, but for periods of concurrent precipitation, intensities would be greater at the higher elevation. In reference to the results from this study, the above concept is unlikely to account for the large differences in intensity. Many of the storms which began or finished with low intensity values had to be eliminated from the study because of instrument accuracy problems. It is during these time frames that Hendrick et al. (1979) attributed much of the decrease in mean storm intensity. To more adequately explain the reasons for variations in intensity, it would be necessary to employ a precipitation recording system which could more accurately discern the exact timing and magnitude of low intensity precipitation. Information concerning cloud structure during storms would also be of assistance.

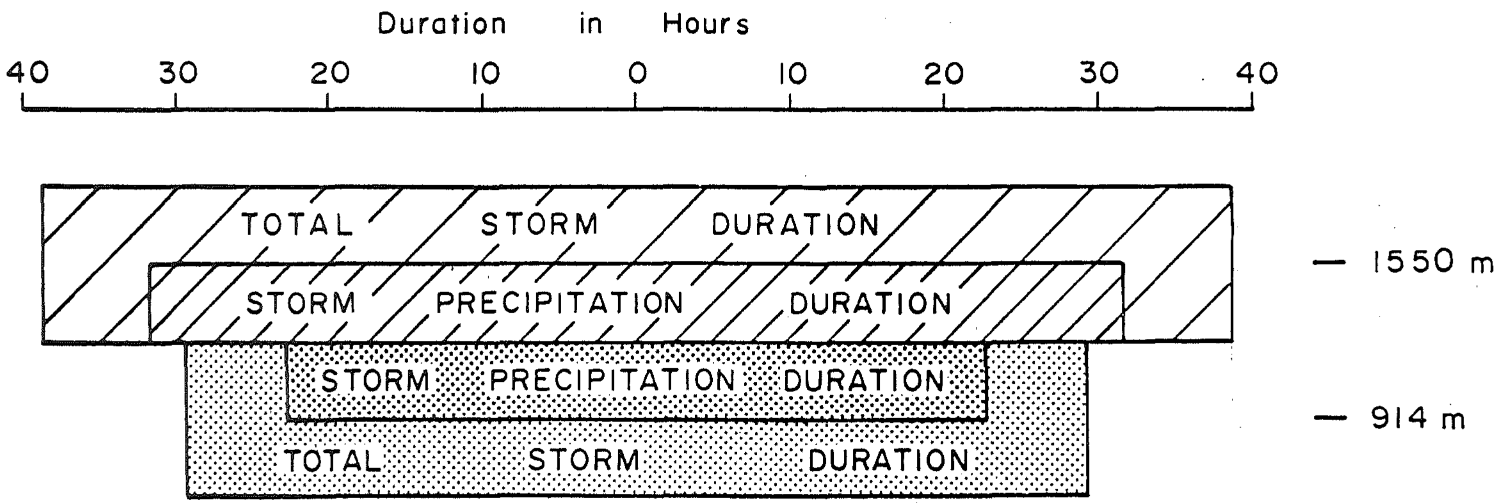


Figure 3.6 COMPARISON OF TOTAL STORM DURATION AND STORM PRECIPITATION DURATION FOR THE SB AND CF CLIMATE STATIONS.

VARIABLES USED IN REGRESSION TO ACCOUNT FOR DIFFERENCES IN PRECIPITATION BETWEEN CF AND SB	% EXPLANATION
A) DIFFERENCE IN STORM PRECIPITATION DURATION	36
B) DIFFERENCE IN STORM DURATION	19
C) DIFFERENCE IN STORM INTENSITY	9
D) DIFFERENCE IN STORM PRECIPITATION INTENSITY	6
MULTIPLE REGRESSION USING B) AND C) ABOVE	43
MULTIPLE REGRESSION USING A) AND D) ABOVE	61

Table 3.5 COEFFICIENTS OF DETERMINATION FOR DIFFERENCES IN STORM INTENSITY AND DURATION CORRELATED WITH DIFFERENCES IN STORM PRECIPITATION BETWEEN THE SB AND CF CLIMATE STATIONS.

In an attempt to more accurately define the importance of storm duration in producing greater quantities of precipitation at higher elevations, regression was performed using inter-storm differences of the storm characteristics as the data base. Differences in intensity were also included and used in a multiple regression with duration. The coefficients of determination from all the regressions are listed in table 3.5 and data plots with the regression equations are contained in appendix D.

The results indicate that it is differences in the duration of precipitation within individual storms which is largely responsible (coefficient of variation = 0.36) for the greater storm precipitation at the SB station. Differences in storm intensity and storm precipitation intensity were found to be relatively unimportant, accounting for only 9 and 6 percent of the variation respectively. However, the combined effects of storm precipitation duration and intensity were revealed in the multiple regression results to account for 61 percent of the variation in the differences in total storm precipitation between CF and SB. The remaining variation may be attributable to measurement and instrument accuracy.

In summary, the main reason for the greater storm precipitation with elevation is the longer duration of actual precipitation during events at the higher elevation. Similarities in totals of storm precipitation between the two elevations would suggest that much of the precipitation may not be derived from orographic or frontal uplift but may be due more to vorticity within cyclonic depressions.

3.6 MEAN FREEZING LEVELS - SNOW STORM TEMPERATURES

Fitzharris (1972) has noted that New Zealand is generally characterized by a mid-latitude west coast climate where the freezing level intersects the alpine zone. Similar conditions are found in the maritime climates of the disturbed westerlies in British Columbia, Washington, Oregon and Chile. Such areas are distinct from the colder continental zones where freezing levels for a majority of the winter are well below the base of the mountains. A summary of freezing levels in Canada by Titus (1968) indicates that, excluding the south-west of British Columbia, the freezing level for over 75% of the days during the winter months is found at the ground surface. Similar conditions are known to exist in other continental climates.

The presence of relatively high freezing levels in the central portions of the South Island is evident in the upper air summaries for Christchurch presented by Tomlinson (1975). Over the period 1958 to 1973, he found the mean monthly freezing level fluctuated from a minimum of 1649 m in July to a maximum of 3203 m in January. He also remarked that typically southerly winds are associated with low freezing levels and northerly ones are associated with high freezing levels. To date no direct evidence has been published to support this. Because of the importance of storm direction in determining snowfall characteristics, a freezing level-wind direction relationship was derived. Based on a 5 year daily record, 1975-1979, of radiosonde readings at Christchurch, figure 3.7 was constructed. The projected surface illustrates the mean monthly freezing level for every 30^{o} difference in wind direction. The relative low number of readings for winds from the easterly direction, especially during the spring months, explains the convoluted nature of the surface in that area. The most important features of figure 3.7, relevant to this study, are the strong contrasts between the elevation of freezing levels from the southerly versus the more

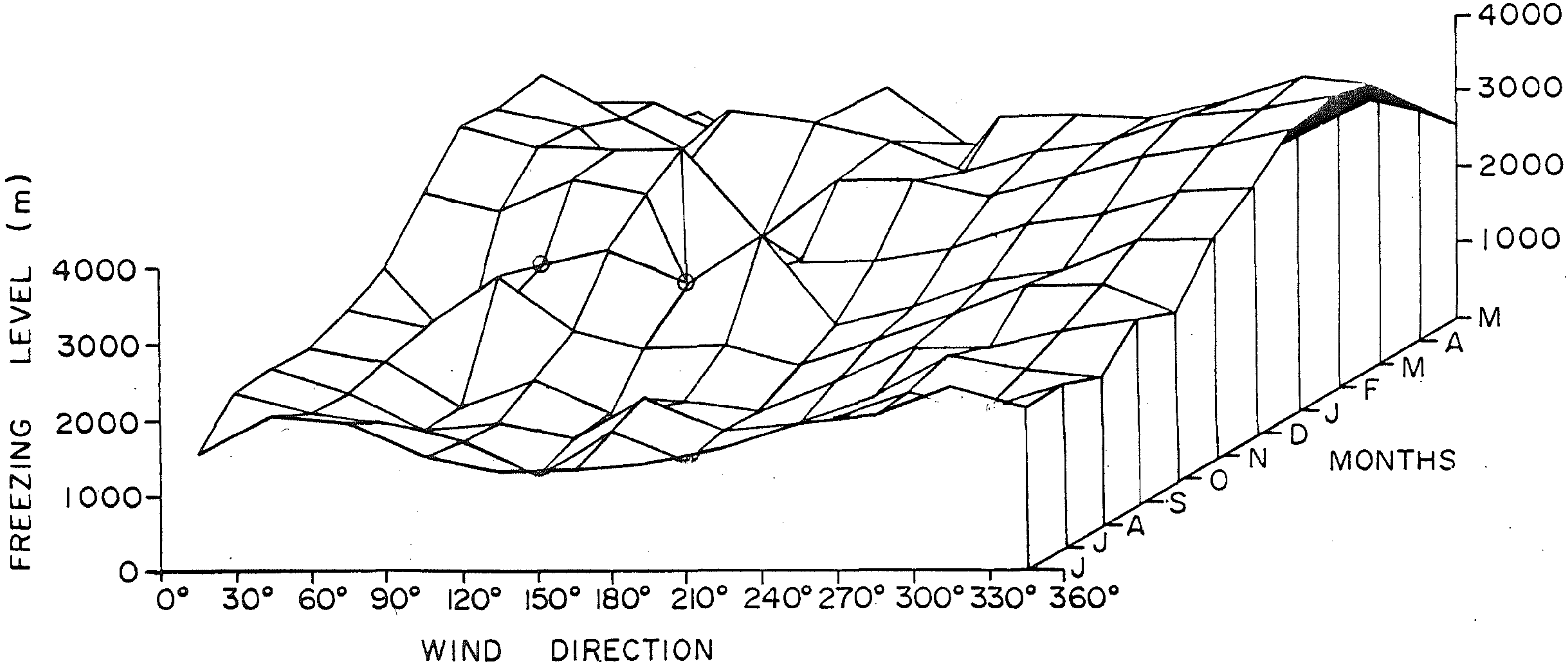


Figure 3.7 MEAN MONTHLY FREEZING LEVEL AT CHRISTCHURCH ACCORDING TO WIND DIRECTION.

northerly, and in particular the north-westerly, wind directions. During the summer months January to March, the freezing level from the south-west averages 3067 m while that from the north-west is over one half kilometer higher at 3627 m. In the winter months of June to August, the freezing level from both directions decreases by just over 1500 m and the difference between the two becomes 614 m. The minimum freezing level from the south-west occurs in August at 1366 m while that for the north-west occurs in July at 1867 m. Although considerable warming of large scale air masses, and hence alterations to the freezing levels, may occur between the alpine regions and Christchurch, these results are indicative of the large differences between, and consistent nature of, freezing levels according to wind direction which occur in the Craigieburn Range.

Within the Craigieburn Mountains, most snow storms are observed to have a normal trend of temperature change, beginning warm and then cooling down. This trend is in general agreement with the passage of a meridional trough of low pressure as described in chapter two and the results of figure 3.7. The warm air at the beginning of the storm commonly belongs to a north to north-westerly airflow preceding the trough and frontal edge. It is subsequently replaced by cooler southerly winds behind the front. If precipitation occurs in all sections of this idealized storm passage, lighter density snow from the cold southerly system should overly the denser and possibly moisture snow generated from the north to north-westerly directions. In terms of avalanche generation, this type of storm produces a much more stable snowpack than that which results from reversing storms. By definition, reversing storms are characterized by relatively cold temperatures at the onset and are followed by a gradual warming. This situation tends to deposit dense heavy snow over lighter density snow which may collapse and avalanche under the weight of the overlying surface.

The presence of winter freezing levels at a comparable elevation to that of the SB meteorological station (1555 m) would suggest that most winter precipitation in the Craigieburn Range should occur in near freezing air temperatures. This view is supported by the frequency of rain and sleet events noted in section 3.3 and by observations made by O'Loughlin (1969a) concerning mean storm temperatures.

The purpose of this section is to detail the range and characteristics of air temperatures associated with the eighty-four storms outlined in section 3.4. It should be remembered that these events were already selected according to a temperature selection technique, hence, any snowfalls which may have occurred at an air temperature primarily above $0^{o}C$ are not included in this analysis. However, as outlined the frequency and magnitude of these storms are thought to be relatively low because of the temperature depression associated with snowfalls of any appreciable quantity.

Analysis of the temperature trends at the SB station for all eighty-four events revealed that only fifty percent remained entirely below the freezing point. Considering the full range of storm temperatures, during even brief non-precipitation periods, seventy percent of the storms recorded maximum temperatures above freezing.

Figure 3.8 illustrates the distribution of these fifty-nine storms throughout the year. During the months October to April all storms recorded above freezing maximum temperatures. The thirty percent of storms which recorded consistent sub-freezing air temperatures are spread through the remainder of the year with a maximum number occurring in July and August (six per month). Notably, however, storms characterized by a maximum temperature above freezing even dominate the mid-winter months.

Fluctuations of storm temperatures about the freezing mark point not only to the high probability of many events containing precipitation in the form of sleet and rain, but also to the potential for surface melting. Fitzharris (1976) has also noted fluctuations in the elevation of the atmospheric melting layer during storms at Mount Cook and has outlined the implications of such for avalanche production.

Within most maritime climates the presence of periods of above freezing temperatures during primarily snowfall events usually implies relatively high minimum storm temperatures. However, in some cases such as reported by Akkourstov (1966) in the Khibiny Range, warm snowfall is frequently followed by rapid temperature decreases of up to 30^{o}C. Although this extreme in temperatures is known not to exist in the Craigieburn Range, it was decided to evaluate the magnitude of minimum storm temperatures which do occur.

Figure 3.8 illustrates the monthly absolute minimum, absolute maximum and mean, minimum storm temperatures for

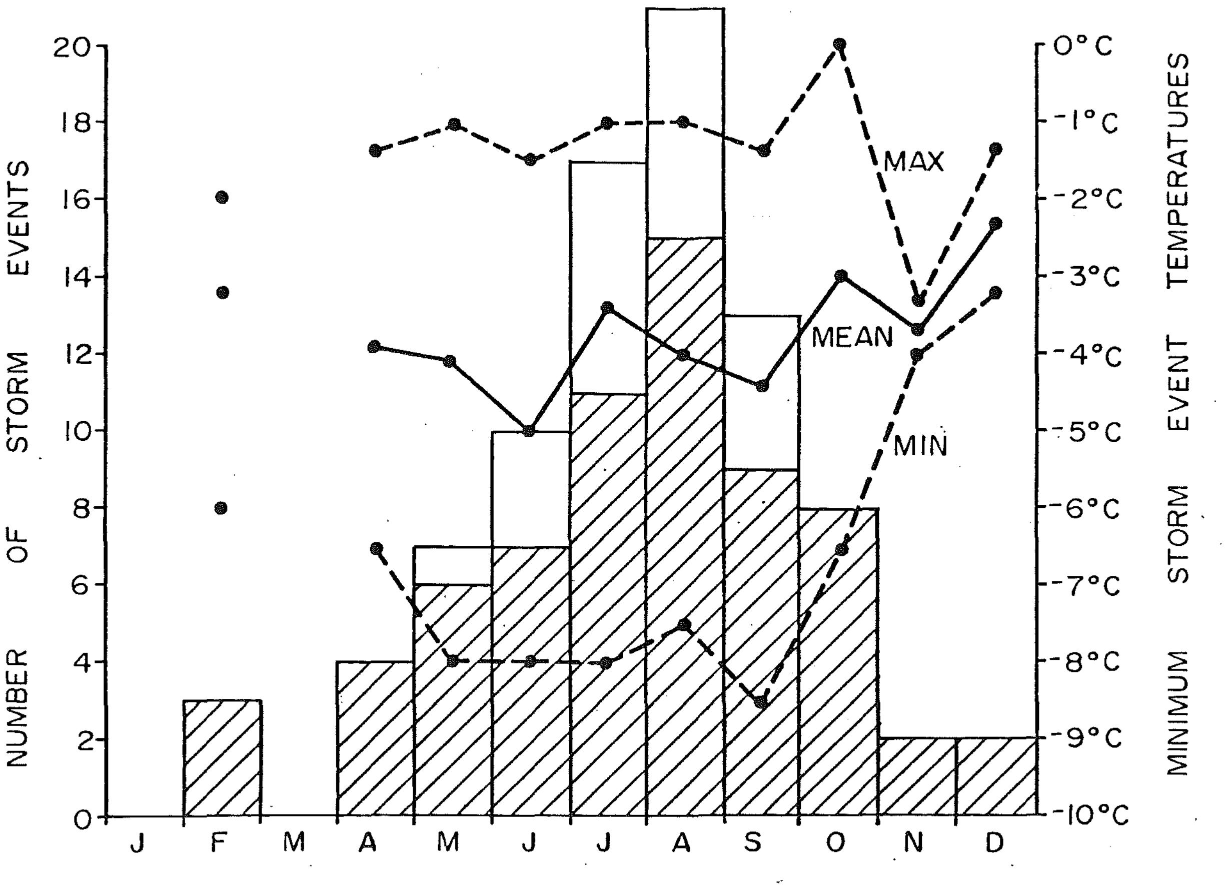


Figure 3.8 MONTHLY FREQUENCY AND TEMPERATURES OF SNOW STORM EVENTS. Lines refer to maximum, mean and minimum monthly minimum storm temperatures. Shaded area of histogram denotes frequency of storms with above freezing maximum temperatures and the clear area those storms which remained entirely below freezing.

the sample of eighty-four events. While there is a weak seasonal trend to the monthly mean minimum temperatures, they tend to remain within a small range between 2 and 5°C below freezing (overall mean = -3.5°C). The coldest storm temperatures are evident in the absolute minimum values which show a rapid increase after the month of September. However, even these values are not exceedingly low and are also relatively infrequent. Only eleven events recorded any temperatures below -7°C and none were less than -9°C.

The small range and relatively warm air temperatures associated with winter precipitation events in the Craigieburn Range are typical of a maritime climate. Similar conditions, however, have also been noted for lower mid-latitude inland mountain locations. For example La Chapelle and Armstrong (1977) report that the air temperature during snowfalls in the San Juan Mountains (3000 m) of Colorado usually range between 0°C and -5°C. Much colder temperatures are known to exist slightly further inland within the Rocky Mountains (La Chapelle 1966) and in the higher latitudes of the continental areas.

3.7 SNOW STORM FREEZING LEVELS

Consistent with fluctuations in the near freezing storm temperatures at the SB site, the position of storm freezing levels in the Craigieburn Range would be expected to cover a range of elevations. This section attempts to define temporal patterns in the freezing levels for the selected eighty-four storms. For each storm, freezing levels were defined at the time of two readily identifiable extremes in snowfall characteristics. These included: the time of maximum precipitation (greatest magnitude and intensity) and minimum air temperature. Temperatures in dry periods within storm events were excluded from this analysis.

The long term mean of the freezing level at the time of maximum precipitation, F_P(m), would be expected, during periods of accumulation and minimal melt, to approximately coincide with the mean snow line elevation. The freezing levels associated with minimum air temperatures, F_T(m), would mark the lower extent of appreciable snow accumulation. Crowe (1971) notes that snow may reach the ground with the freezing level at a maximum of 300 m above the surface. At greater elevations, the snow crystals would be entirely melted or evaporated before reaching the surface. Also, heat consumed in melting or evaporating the falling snow will tend to cause a progressive downward cooling and lowering of the freezing level (Lumb 1961). Hence, only minimal snow deposition is likely to occur below the elevation of the lowest freezing level during any given storm.

In order to calculate the freezing levels for the various storms from thermograph records, it was necessary first to calculate environmental lapse rates. Sufficient reliable information from the temperature profile was available to calculate the two types of freezing levels for eighty-one storm events, the mean lapse rate for which was calculated to be 0.62°C per 100 m (maximum error of estimate

= $\pm$ 0.04 at a 95% confidence level). This storm environmental lapse rate is very similar to the moist adiabatic lapse rate of 0.58°C per 100 m suggested by List (1966) for air temperatures between 0°C and -5°C at 850 mb, which is the approximate pressure level corresponding to the SB meteorological station. It is also in good agreement with the mean lapse rate of 0.6°C per 100 m suggested by Obled and Harder (1979) for mountainous regions.

In the final calculation of mean freezing levels, some additional storms had to be excluded from the eighty-one storm sample because extrapolation placed the freezing level well beyond the range of elevations in which the lapse rates were originally calculated. The final mean monthly freezing levels, pictured in figure 3.9, were therefore calculated on a total of eighty storms at the time of maximum precipitation and seventy-six at the time of minimum air temperature. The mean freezing levels appear to be good estimates according to the confidence limits (at 95%) which do not vary more than $\pm$22 m for F_P and $\pm$34 m for F_T.

The seasonal trend in the elevation of F_P is relatively small, varying from February to August by only 300 m (no data were available for January or March). The F_P during the main winter months lies between approximately 1100 and 1300 m with the lowest value being recorded in August. These elevations closely correspond to the snow lines observed during the 1978-80 field seasons and to similar observations made by O'Loughlin (1969a). They are also only 600 - 800 m lower than the mean permanent snowline suggested by Chinn (1975) for the Waimakariri Basin. A

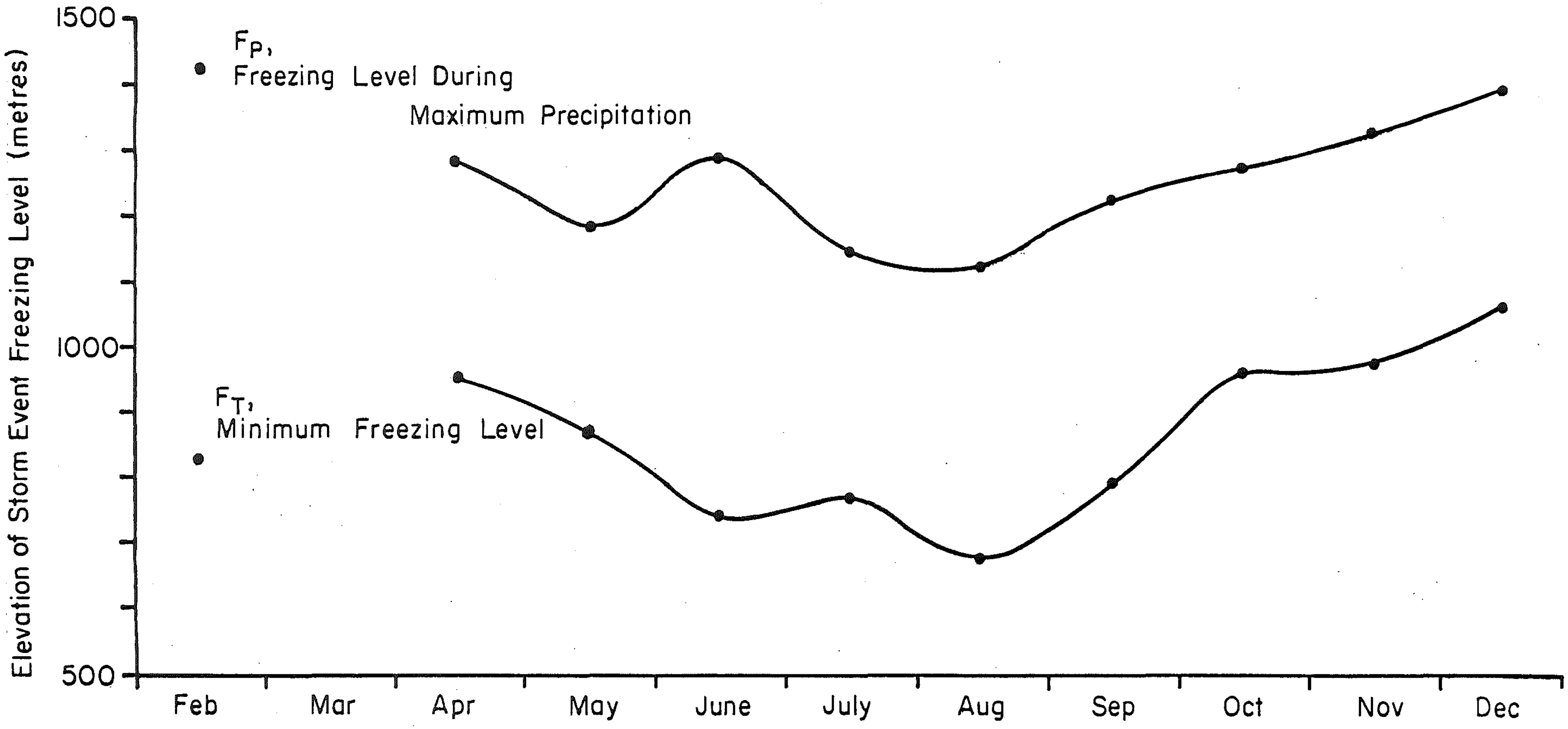


Figure 3.9 MEAN MONTHLY ELEVATIONS OF SNOW STORM FREEZING LEVELS.

greater variation exists in the F_T values, than found for F_P, ranging from 1060 m in December to 680 m in August. Consistent with the elevations for the winter F_T values (approximately 700-800 m), snowfall is regularly recorded at the CF climate station (914 m) and observed at lower elevations. In most cases, these snowfalls are short-lived and usually produce a minimal snow coverage. The major snow accumulations occur above the treeline, which is only $\pm$100 m different from the F_P elevations.

3.8 SYNOPTIC PATTERNS OF LARGE SNOWFALLS

The meteorology of snow producing storms in New Zealand is highly complex such that extensive data input is required for in-depth analysis. Within the alpine areas where meteorological data is scarce, most analyses of snowfall have primarily involved a simple classification of surface pressure patterns. O'Loughlin (1969a) classified thirty snow storms in the Craigieburn Mountains into two broad synoptic types; the first was related to the normal passage of a trough of low pressure over the country, as described in section 2.3.1, and the second involved a deepening depression tracking to the east of New Zealand accompanied by a cold southerly airstream over the South Island. Over 75% of the thirty storms were found to be of the first type.

Based on a slightly more elaborate synoptic classification, Owens and Prowse (1980) have shown that total winter snowfall at Coronet Peak is related to the winter frequency of south-west airflow following the passage of cold fronts.

A more detailed meteorological analysis of ten snowfalls, which caused disruption to the lowlands of the South Island between 1967 and 1975, has been conducted by Neale and Thompson (1977). Based on the results of Goree and Younkin (1966) in the United States, they classified two major types of snow storms: warm advection and cold vorticity. Because of the lack of conventional meteorological data, subjective interpretations had to be made of many of the storm conditions. Generally, Neale and Thompson (1977) considered storms to be of the warm advection type when the structure of the atmosphere was similar to that of a warm front as found in "Pacific northwest" storms noted by Younkin (1968). In such storms, heavy snowfalls are generated when moist warm air is uplifted, along the warm or quasi-stationary front and overtop colder air in the lower troposphere. For cases where warm advection was limited and large scale vorticity advection prevailed, in broad flows of unstable cold air originating from high latitudes, the classification of cold vorticity was employed. Such conditions are typical of 'polar low' storms which are important over Britain (Lyall 1972).

An in-depth classification of snow storms similar to that by Neale and Thompson (1977) is beyond the scope and objectives of this report. However, some insight concerning snowfall characteristics can be gained through the following comparison of synoptic patterns associated with the major snow storms which occurred during the course of this study.

Appendix E contains the precipitation and wind data (recorded at the SB and CF stations respectively) for the fifteen storms cited in section 3.4 as having produced more than 25 mm of precipitation in the form of snow. These fifteen storms occurred over a six month range from May to October, although over fifty percent were concentrated in the month of August. The surface, 850, 700, 500 mb and 1000-500 mb thickness charts were reviewed for each storm and the surface charts corresponding to the time of maximum precipitation in each storm are also presented in appendix E.

Surprisingly, only three of the storm pressure patterns (June 9, 1976; August 17, 1978; August 13, 1979) could be considered typical of the normal passage of a trough of low pressure over the country as described in section 2.3.1. The storm of June 8-10, 1976 exemplifies this type of storm in which a meridional trough of low pressure and accompanying cold front extended from a deep depression located to the south-west of the country. Strong north-westerly winds in excess of 20 knots preceded the front and the major snowfall period by over twenty-four hours. As the cold front passed over the Craigieburn Mountains early on June 10, precipitation intensified and the winds shifted to a southerly direction. Similar but less distinct trends occurred over the periods August 15-18, 1978 and August 12-15, 1979.

For the remaining twelve storms, cold fronts did not appear to be as significant a feature as the fact that the centres of low pressure within the trough were located either directly or in close proximity to the centre of the South Island. In almost all cases, these surface depressions were found to be associated with centres of cyclonic vorticity at the downstream edge of short wave troughs at the 500 mb level. According to Goree and Younkin (1966), snowfall may extend over broad areas around these systems, frequently 500 - 1000 km ahead and 500 km poleward of the 500 mb centres of cyclonic vorticity. In this study, it was found that snowfall frequently began while the depressions were still off the western coast of the South Island. Westerly and north-westerly winds usually prevailed at this stage just as in conditions preceding the passage of a cold front. As the depression passed over or close to the centre of the South Island, wind directions changed considerably and precipitation usually intensified. In cases where the depression passed off to the east coast of New Zealand, snowfall continued at the SB station primarily from the north-east to south-east quarter. The storm of September 13, 1976 is the best example of this situation.

The eastward progression of centres of cyclonic vorticity are normally rapid except when slowed by the presence of large blocking anticyclones. Browne (1975) has found this type of situation not uncommon near New Zealand, and most prevalent in autumn and winter with anticyclones to the south-east of the country. The extent to which this affected some of the storms listed in appendix E is unclear, but is suspected in a number of cases where lows tended to stall off the east coast.

Generally, all the storms except those which were distinctly the result of cold front passage were characterized by synoptic patterns similar to those of cold vorticity storms identified by Neale and Thompson (1977). However, in a few cases (August 19, 1975; August 25, 1975; June 28, 1977; October 19, 1978; August 13, 1979) warm or quasi-stationary fronts, typical of warm advection storms, were found in association with the centres of cyclonic vorticity. An evaluation of the relative importance of warm advection along the frontal surfaces versus vorticity advection in producing ascending motion and subsequent snowfall at the Craigieburn Mountains could only be obtained by a thorough examination of meteorological conditions. In view of the scarcity of mountain climate data and the complicating effects of alpine topography, as in blocking and/or trapping cold air, such an analysis would be expected to be more speculative than that performed by Neale and Thompson (1977) for the lowland snowfalls.

The major point to be made from this analysis is that most major snowfalls in the Craigieburn Range over the period 1975-80 did not occur from the simple passage of a cold front. Instead, they were associated with surface depressions and upper level centres of cyclonic vorticity which tracked directly over or in close proximity to the centre of the South Island. This is consistent with Thompson and Kells' (1973) report that during the period 1964-1969 low pressure systems were responsible for 50% of the days with snowfall on Mt Ruapehu. The importance of cyclonic vorticity also supports the results in section 3.5 that precipitation at the SB and CF sites is related more to vorticity than orographic or frontal uplifting.

3.9 NEW SNOW DENSITY

Snow falling from the atmosphere is known to accumulate on the ground with a density primarily dependent on the original ice crystal forms and the conditions under which they were deposited. Over time the density of snow will change as the result of a number of processes described in chapter four. Seligman (1936) and similarly Paterson (1969) have devised a classification of density ranges according to the type of snow (table 3.6). Only the first six rows of table 3.6 apply to snow originating directly from the atmosphere.

A number of researchers have made observations of snow densities in New Zealand (table 3.7) but for the most part their results are not comparable nor representative of new snow conditions in alpine areas. The observations include samples from wind packed snow, snow-sleet storms, sub-alpine locations and some are based on only single events. This report attempts to properly classify new snow densities for the Craigieburn Range so that comparisons may be made to other regions. Within this particular section, only snow deposited under relatively calm conditions is considered. Snow which was noticeably wind compacted is discussed in conjunction with post-depositional processes in chapter four. Mixed rain and/or sleet events, which tended to produce densities in excess of 400 kg m^{-3}, are also excluded from this section.

A total of forty-one density measurements were made during the 1978, 1979 and 1980 winters of new snow deposited in calm conditions. Any snow not sampled directly after

TYPE OF SNOW-ICE	DENSITY $kg\ m^{-3}$
WILD SNOW	10-30
NEW SNOW FALLING IN CALM CONDITIONS	50-65
DAMP NEW SNOW IMMEDIATELY AFTER FALLING	100-200
VERY SLIGHTLY WIND TOUGHENED SNOW, IMMEDIATELY AFTER FALLING	63-80
AVERAGE WIND TOUGHENED SNOW	280
HARD WIND SLAB	350
SETTLING SNOW	70-190
SETTLED SNOW	200-300
NEW FIRN SNOW	400-550
ADVANCED FIRN SNOW	550-650
THAWED FIRN SNOW	600-700
FIRN-ICE, VERY WET SNOW	800
GLACIER ICE	917
WATER	1000

(AFTER SELIGMAN 1936)

Table 3.6 CLASSIFICATION OF SNOW AND ICE FORMS ACCORDING TO DENSITY.

RESEARCHER	NEW SNOW DENSITY ($kg\ m^{-3}$)	LOCATION	COMMENTS
HEINE (1962)	300-500	MT RUAPEHU	SURFACE WIND PACKED SNOW
GILLIES (1964)	100-500 EXTREMES; 200-700 NORMAL RANGE; 300 MEAN.	FRASER CATCHMENT, CENTRAL OTAGO.	1962 WINTER
MORRIS AND O'LOUGHLIN (1965)	110-200 150 MEAN	CRAIGIEBURN MOUNTAINS	1962-1964 WINTERS
CHINN (1968)	400-500 200-250	SOUTH CANTERBURY FOOTHILLS INLAND SOUTH CANTERBURY	NOVEMBER 1967 SNOWFALL "
HUGHES (1969)	210-240	CRAIGIEBURN MOUNTAINS	NOVEMBER 1967 SNOWFALL
O'LOUGHLIN (1969a)	80-280 160 MEAN	CRAIGIEBURN MOUNTAINS (APPROXIMATELY 1000 m ELEVATION)	1964-1968 SEVEN STORMS
HUGHES (1974)	300-400 80-100	SOUTH CANTERBURY LOW ELEVATIONS HIGH ELEVATIONS	AUGUST 1973 SNOWFALL "
WEIR & OWENS (1981)	91-486 179 MEDIAN	MT HUTT 1600 m	1979 WINTER SEVENTEEN STORMS
THIS STUDY	60-220 130 MEAN	CRAIGIEBURN MOUNTAINS	1978-1980 WINTERS CALM CONDITIONS

Table 3.7 MEASUREMENTS OF NEW SNOW DENSITY IN NEW ZEALAND.

initial deposition and which was considered to have experienced some initial equi-temperature metamorphism (described in chapter four) was not included in this analysis. The distribution of the new snow densities appears in figure 3.10.

No observations were made of what Seligman (1936) termed wild snow, ranging in density from 10 - 30 kg m^{-3}. Such very low values are not considered to be exceptional in Scandinavia (Morris and O'Loughlin 1965) but are even less than those noted to be of low density (mean 85 kg m^{-3}) within the Colorado Rocky Mountains (Grant and Rhea 1974). Thirty percent of the samples were found to be less than 100 kg m^{-3} and contained one observation of 60 kg m^{-3} which is the lowest ever reported within New Zealand. However, a majority of the new snow densities fall into the range considered by Seligman (1936) to be typical of damp new snow. The total range of densities has a mean value of 130 kg m^{-3} (median = 123 kg m^{-3}), one which is not dissimilar to that observed for some cold continental mountains and tundra-Arctic climates [Bilello (1969); Steppuhn and Dyke (1974)].

As an aid to weather and avalanche forecasting, the density of new snow has been compared to both upper air and surface air temperatures and for many climates has produced quite good correlations [for example: Bossclasco (1954); Diamond and Lowry (1954)]. Attempts were made to correlate the density of new snow from the Craigieburn Range with the 700 and 800 mb temperatures recorded from upper air soundings at Christchurch but only weak correlation coefficients of 0.33 and 0.46 resulted (appendix F).

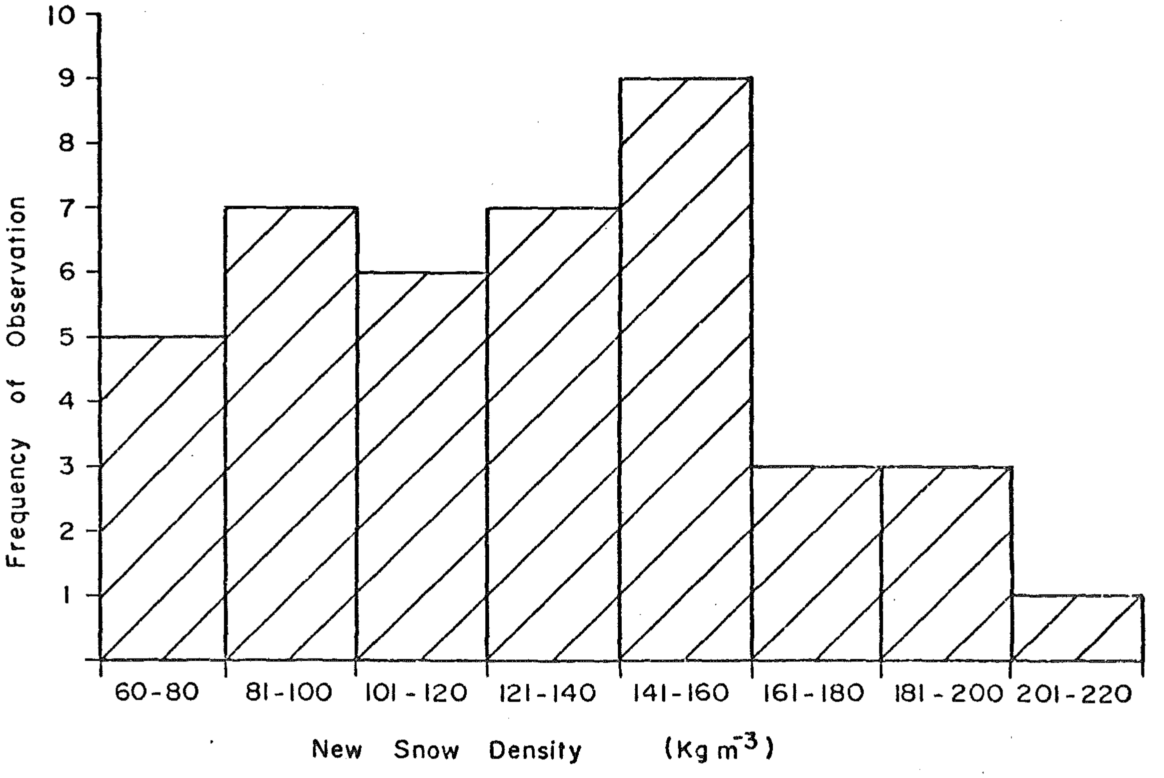


Figure 3.10 FREQUENCY HISTOGRAM OF NEW SNOW DENSITY. Deposited under calm conditions.

A much higher correlation coefficient of 0.71 was found to exist between the surface air temperatures and new snow density (figure 3.11). The range of air temperatures used in the regression analysis are typical of storm temperatures described in section 3.6, with a mean of 2.9°C below freezing and only one value less than -6°C. Three density values were recorded with air temperatures slightly above freezing (all less than 1°C) but all recorded densities greater than approximately 150 kg m^{-3}.

The density of new snow has also been suggested by Nakaya (1951) and Gold and Powers (1952) to be a function of the size and shape of snow crystals which is in turn related to the temperature and degree of supersaturation at the time of formation. In general, needle crystals form predominantly above -8°C, plates and dendrites between -8°C and -20°C and columns at less than -20°C. Although no accurate frequency record of crystal types was made during this study, needle and stellar crystals were observed on a majority of occasions. The presence of needle crystals is consistent with La Chapelle's (1969) observation that these crystals are most prevalent when surface air temperatures are near the freezing mark. It might be expected that because of the shape of needle crystals, they would easily compact and form relatively high density snow. However, needle crystals were observed at both extremes of the density range in figure 3.11. This may largely be explained by the presence of riming. Powers et al. (1964) found that the new snow density for needle crystals may be increased by 30 percent from moderate riming. Observations from this

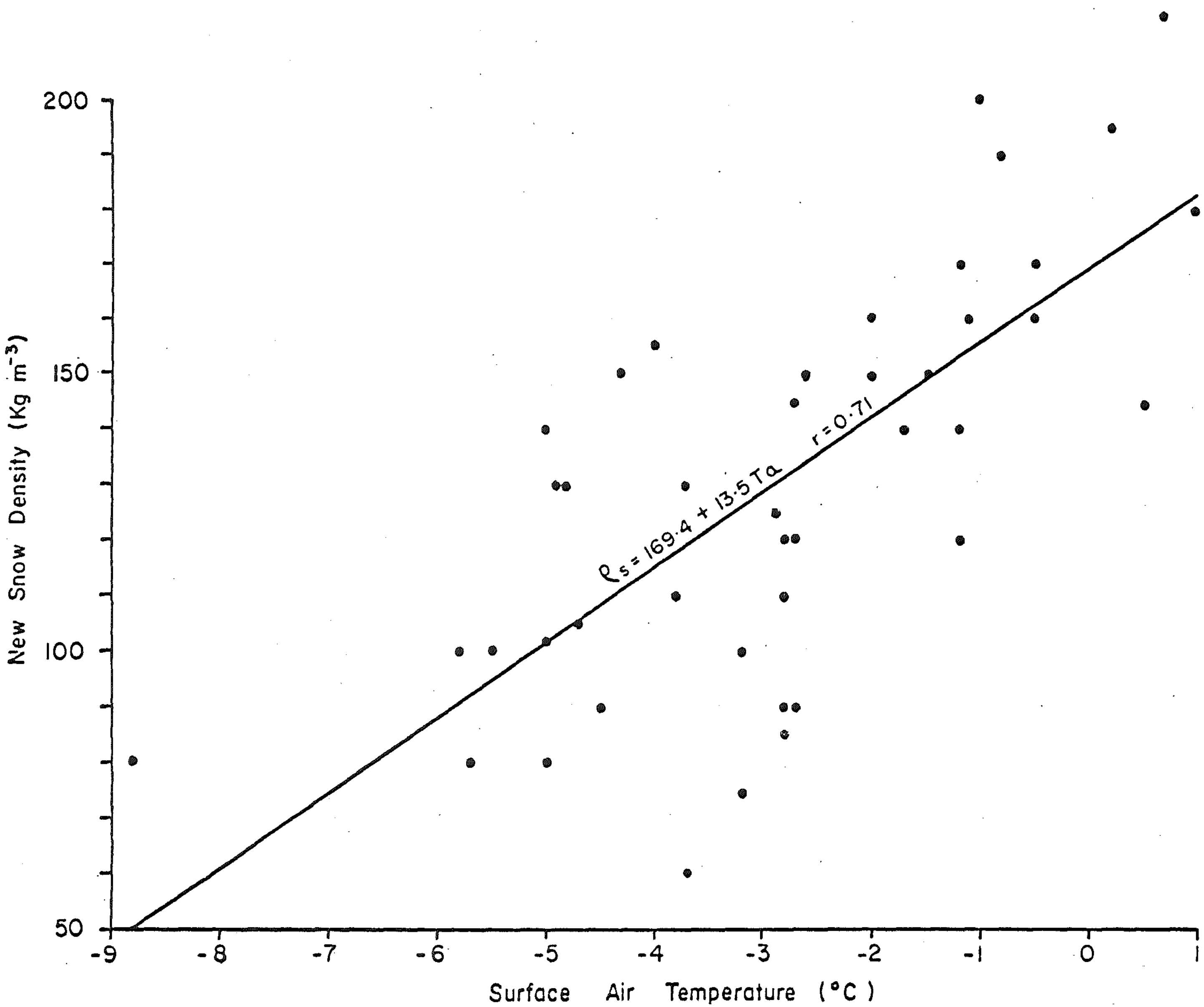


Figure 3.11 REGRESSION OF NEW SNOW DENSITY WITH SURFACE AIR TEMPERATURE.

study revealed that needle crystals, in the presence of riming, tended to bond together forming large snowflakes (figure 3.12), similar in structure to dendritic and stellar crystals which are known to produce low density snow. Some degree of riming was present on a majority of the crystal forms observed in the Craigieburn Range. This is consistent with La Chapelle's (1969) observation that relatively warm air, strong convection and rapid lifting of moist air by steep mountains - meteorological conditions similar to those observed for the Southern Alps - favours crystal riming. On numerous occasions heavily rimed germs (immature crystal forms) were observed, indicating that the elevation of the clouds producing precipitation were relatively low. The heavily rimed large graupel form was observed only rarely.

3.10 SUMMARY

Snowfall is a significant component of yearly precipitation above 1500 m in the Craigieburn Range with the major contribution occurring between May and October, although, some snow may fall, even to low elevations, at any time of the year. Similarly, rain regularly occurs even in the middle of winter but the smallest contributions are found in August when snowfall is usually maximized.

Seventy-five percent of the snow storms were found to produce less than 25 mm of total precipitation and were characterized by intensities which for 75% of the total

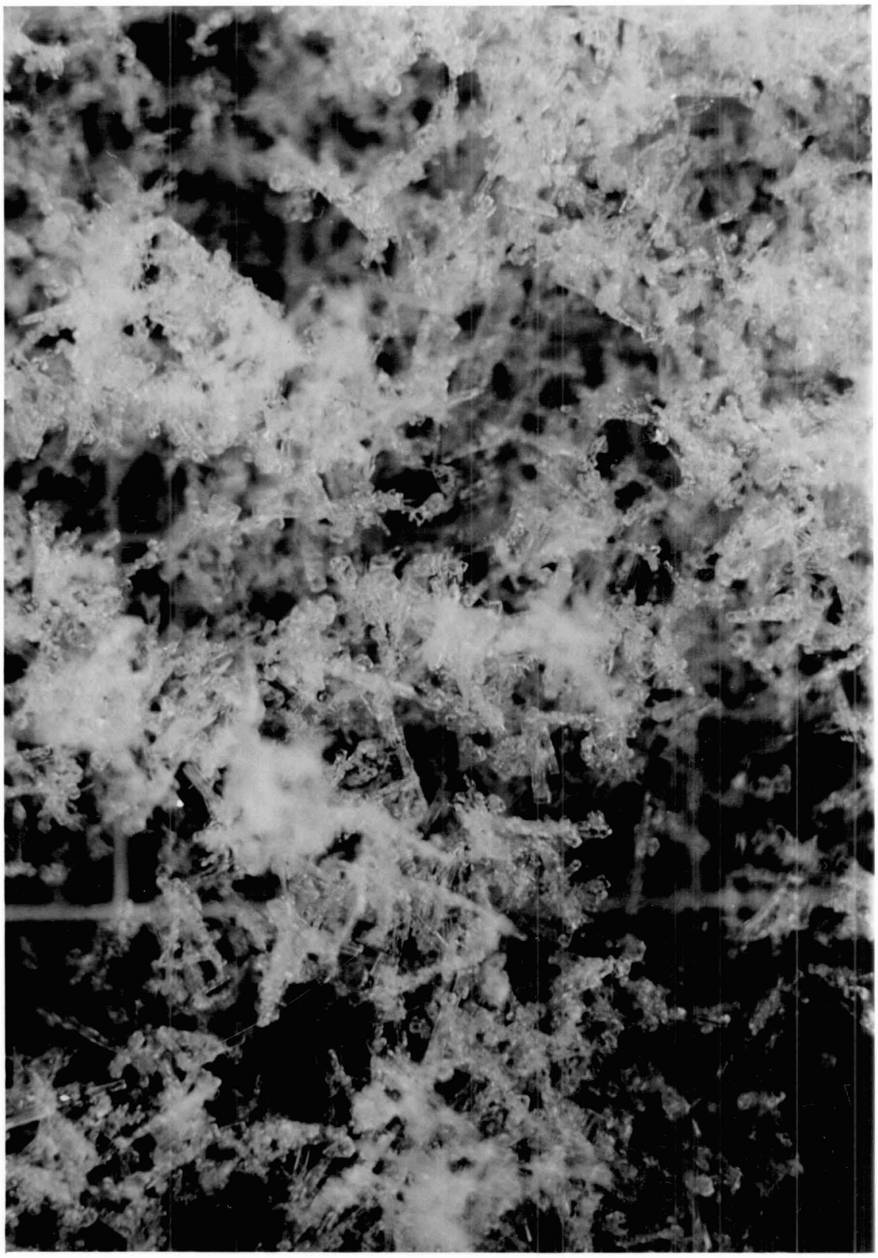


Figure 3.12 RIMED NEEDLE CRYSTALS. The presence of rime tends to bond the needle crystals together forming snowflakes similar in structure to multi-pointed crystals such as dendrites or stellar crystals. These types of crystals when deposited have a high void ratio and produce light density snow. When riming is extensive on individual crystals a much higher density snow is found, especially in the presence of strong winds.

snowfall were less than 2.5 mm hr^{-1}. However, an average of one storm per year produced more than 25 mm of snow at a rate greater than 2.5 mm hr^{-1}. Such conditions according to overseas experience are critical for the production of direct action avalanching.

Inter-storm snowfall was observed to significantly increase with elevation but was not due to increases in snowfall intensity which in fact seemed to decrease with elevation. This suggested that precipitation was not directly dependent on orographic or frontal uplifting. Longer storm duration at the SB compared to the CF station offered the best explanation for increases in precipitation with elevation.

Most major snowfalls originated from centres of cyclonic vorticity which passed over or in close proximity to the centre of the South Island. The passage of cold fronts with the associated depression at some distance south of the country are much less important in producing heavy precipitation. In both storm situations, the onset of precipitation was usually accompanied by a west to north-westerly airflow. Southerly winds tended to precede the passage of cold fronts but a variety of wind directions, depending on the particular track of the centre of vorticity, characterized the passage of depressions. Winds from an easterly quarter were usually associated with depressions which passed to the eastern coast of the country.

According to Christchurch radiosonde data, the highest mean freezing levels are related to west to north-westerly wind directions and the lowest with more southerly winds. The average winter freezing level approximately intersects the 1500 m elevation of the SB climate station. Overall storm temperatures at the SB site were consistently near the freezing mark with 70% of the storms containing some above freezing temperatures. Minimum storm temperatures were also relatively warm, averaging only -2 to -3^{o}C and at a maximum rarely falling below -6^{o}C.

The mean monthly freezing level for periods of maximum storm precipitation ranged over the winter from 1300 m in June to 1100 m in August. However, minimum mean monthly storm freezing levels were found to be approximately 450 m lower.

Snowfall originating under calm conditions has a mean density of 130 kg m^{-3}, similar to values reported for some continental climates. Riming, an important part of the environment, was found to be responsible for the production of both high and low density snow.

CHAPTER IV

SNOWPACK STRUCTURE

4.1 INTRODUCTION

A range of changes occur in the structure of snow once it has been deposited on the ground. Over a relatively short period of time the ice matrix bears little resemblance to the crystal forms which originated in the atmosphere. The rate and nature of the changes which take place are directly related to the climate in which the snow is deposited. Hence, for a given region the snow structure will reflect the climate of that area.

The main objective of this chapter is to define the snow structure of the Craigieburn snowpack with respect to the local climate. The chapter is broken into six main sections, the first of which is a summary of research already conducted in New Zealand. The subsequent three sections detail the post-depositional changes which affect snow once on the ground. These have been divided into three areas: wind effects, metamorphic processes by vapour transfer and melt-freeze metamorphism. Each of these sections includes theory on the particular processes involved and results of field observations. The next section considers the temporal fluctuations in the mean snowpack density and attempts to classify the snowpack in relation to overseas results. A brief summary of the results concludes the chapter.

4.2 RESEARCH ON SNOW STRUCTURE IN NEW ZEALAND

Prior to 1980, published information concerning the structure of the New Zealand snowpack was almost non-existent. Heine (1962) provided some basic information regarding snow densities and grain sizes for Mt Ruapehu, and Morris and O'Loughlin (1965) reported on snowmelt densities for the Craigieburn Mountains. However, the amount of snowpack information in these reports was extremely limited. The most extensive work was presented in an internal Forest Research Institute report by O'Loughlin (1969a). Data concerning snow quality, density and temperatures were described for three vertical profiles of the snowpack. Other data have also been collected, primarily by people concerned with the avalanche hazard, but this information is largely disjointed and relatively inaccessible.

The first published work containing detailed information about snowpack structure appeared in 1981 by Owens and Prowse (1980) and soon afterward McNulty and Fitzharris (1980) presented snowpack information concerning an avalanche event in a Canterbury alpine basin. Both of the above reports pointed to the significance of depth hoar crystals which prior to the mid 1970's had generally not been thought to exist in New Zealand. Weir (1979) in an unpublished thesis has also attested to the presence of this particular crystal form in the Mt Hutt region. Much of the work conducted on the snowpack at Mt Hutt has been subsequently published by Weir and Owens (1981). Specific details of the above reports are discussed in the relevant sections of this chapter.

Definitely, there is an apparent scarcity of published material describing the stratigraphy of the New Zealand snowpack and more specifically the importance of the alpine climate in controlling the development of the various snow forms and structure. The following sections are an attempt to improve this situation.

4.3 WIND DEPOSITED SNOW

4.3.1 Background

Wind may affect snow both during initial precipitation and once it has been deposited. The forms and structures of snow which wind produces are distinct from those resulting from snowfall during calm conditions. For this reason, wind deposited snow is discussed in this chapter rather than in chapter three.

In some parts of the world, snow is deposited exclusively under conditions of high wind. For example, Akkouratov (1966) reports that in the Khibiny Range of the U.S.S.R. solid precipitation almost always occurs under blizzard conditions. For most regions snowfalls occur under a range of wind conditions, although wind action subsequent to precipitation may rework all new snow inputs. The ability of wind to transport snow depends on the strength of the wind and the shear strength of the snow surface. Windspeeds as low as 2.4 m s^{-1} have been found for parts of Europe to be sufficient for the redistribution of loosely bonded snow (Kunquerstev 1971) but more generally Mellor (1964) applies a range of 3.0 to 8.0 m s^{-1}. He also mentions that for surface crusts formed by sentering or freeze-thaw metamorphism, winds of up to 30 m s^{-1} are required before surface erosion will occur.

As first outlined by Seligman (1936), wind may produce two broad types of snow structure, wind crust and wind slab. Wind crusts are principally formed on windward slopes where the wind strikes with greatest force. The thickness of these crusts is normally only 1 - 2 mm (Martinelli 1971), much less than the more extensive wind slab layers which usually form on leeward slopes. In both cases, suspended snow is packed into the surface as the forces from impact and wind turbulence destroy the original crystal shapes. New wind compacted snow is usually firmly bonded and is characterized by higher densities and smaller grain sizes than those for snow deposited under calm conditions. Benson (1979) suggests grain sizes for wind packed snow will range from fine, 0.5 - 1.0 mm, to very fine, less than 0.5 mm, depending on wind speed. Similarly, the final density of the snow is a function of the force of the windflow.

The overall mechanical strength of wind packed snow has been found to be well correlated with the degree of inter-granular bonding which is in turn related to density. Seligman (1936) points to the role of moisture laden wind in bonding snow through the condensation of water vapour directly onto the grains as they are compacted. More recently La Chapelle (1969) has found that heavily rimed crystals also assist in the cementing of the snow matrix, and tend to produce quite stiff wind packed layers.

Seligman (1936) differentiated two types of snow slab, hard and soft, according to the strength of windflow and supply of snow in the air. Soft slab was said to form when the amount of precipitated or drifted snow was too great for the wind to solidly pack. As such, it tended to occur over quite large areas. In contrast, hard slab was much more localised and firmly bonded. Considerable research has been conducted attempting to further identify the process responsible for hard slab formation and the general contention [La Chapelle (1966); Judson (1967); Martinelli (1971)] is that cold air temperatures as well as high windspeeds are required during deposition. The presence of cold air temperatures would seem to preclude the possibility of high moisture levels and rimed crystals. However, many of the observations of hard slab have been in cold continental climates and it is possible for more maritime climates that moisture plays a more important role in hard slab formation.

4.3.2 Field Observations in the Craigieburn Range

As described in the review of the alpine climate for the Craigieburn Mountains, high windspeeds frequently dominate the alpine zones. Windspeeds during the winter months average 2.5 to 3.5 m s^{-1} at the SB site, with mean daily maxima in excess of 6.0 m s^{-1}. According to the results of Kunquerstev (1971), these windspeeds are sufficient to ensure the complete redistribution of loosely bonded snow.

These quoted windspeeds are only representative of a mid-elevation site (1550 m). Much greater speeds usually occur near ridge-top locations (La Chapelle 1970). Pressure plate maximum gust recorders have been used for a seven month period at the summit of the Craigieburn Range (1800 m). Rowe (1968) reports that during three of these months the gust meter was pushed to the scale limit of 67 m s^{-1} and gusts exceeding 30 m s^{-1} occurred in all months except February. Similarly, McSaveney (1978) has reported the daily average windspeed for a one year return period on nearby Mt Hutt (2078 m) to be in excess of 50 m s^{-1}. During a north-westerly airflow on Mt Cockayne this researcher also observed that ridge-top winds were sufficient on a leeward slope to remove a crust layer over 50 mm thick in blocks 0.5 to 1.0 m^2 in area. These sections were transported in the air at a height of approximately 10 m over considerable distances. The removal of this protective crust also allowed subsequent wind scour to totally denude the ridge locations of snow. The scale of this deflation process is definitely not considered in the normal saltation and turbulent diffusion models (Budd 1966) developed for blowing snow. The rate and magnitude of snow redistribution during such periods is expected to be immense. Accurate measurements of the redistribution is impractical even with systems specifically designed to measure blowing snow [Schmidt (1977); Tabler and Jairell (1971)].

As wind deposited snow appears to be so significant, at least in the higher elevations of the Craigieburn

Mountains, attempts were made to identify some of the surface snow properties associated with wind deposition. Information concerning snow density, grain size and mechanical strength was extracted from the 1979 and 1980 snow pit records for all surface layers which were known to be wind affected.

The layers selected for analysis contained some snow which had been wind deposited one or two days prior to the snow pit analysis. This length of time was considered insufficient to appreciably alter the original snow properties of the wind packed layers. In some cases, lighter snow had been deposited on the wind compacted snow but the effect of this layer on the underlying snow was also considered negligible. Layers which had undergone appreciable warming or melt-freeze metamorphism were not included for analysis.

A total of forty-two different wind deposited surface layers were identified. The results of the density measurements appear in figure 4.1 and in comparison to the new snow densities described in chapter three evince the role of wind in the densification process. The average density was calculated to be 255 kg m^{-3}, approximately 130 kg m^{-3} higher than the average snow density recorded in the absence of strong winds. A difference of means t-test indicated a significant difference between the two sample sets at a 95% confidence level.

Although strict density criteria for discerning hard and soft wind slab have not been established, Martinelli (1971) suggested that values of greater than

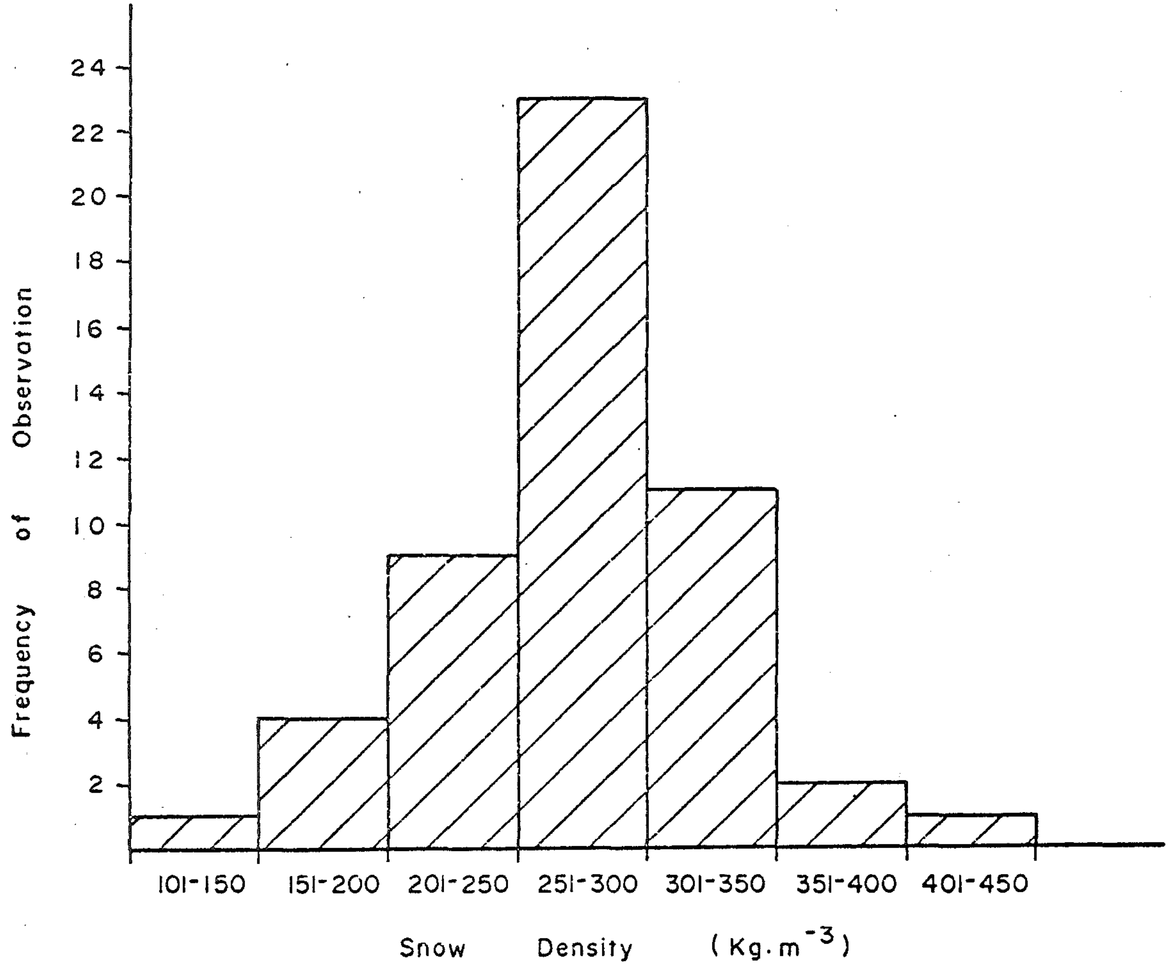


Figure 4.1 DENSITY OF WIND DEPOSITED SNOW. Mean density of forty-two different snow strata identified as being originally deposited under conditions of high wind, mean = 255 kg m^{-3}.

290 kg m^{-3} were characteristic of 'initial hard slab' and Gray (1979) indicated that true hard slab normally has a density of approximately 350 kg m^{-3}. Only three of the forty-two observations exceeded the latter value while 30% were greater than the former.

The individual grain sizes recorded for the above mentioned layers never exceeded 0.5 mm in diameter. This maximum agrees with the category of very fine grains suggested by Benson (1979) for heavily wind packed snow. Martinelli (1971), however, reports that wind packed grains, even in hard slab layers, are significantly larger at between 0.4 and 0.7 mm while Seligman (1936) is more in accordance with Benson's earlier statement describing the grain size for hard slab layers as approximately 0.2 mm. The frequency distribution for all grain sizes of wind deposited snow from this study appears in figure 4.2. The modal class was found to be 0.21 to 0.3 mm, with an overall mean of 0.24 mm. By any classification these sizes must be considered as 'very fine'.

Because of the volume-sphere relationship a positive correlation was expected to exist between increasing density and decreasing grain size. However, an almost null correlation was calculated between the two variables. In plotting the relationship (figure 4.3), a full range of grain sizes was evident for most densities. The most probable explanation for the poor correlation is the errors inherent in measuring the mean grain size of such fine textured layers in a field situation.

In addition to small grain size and high density,

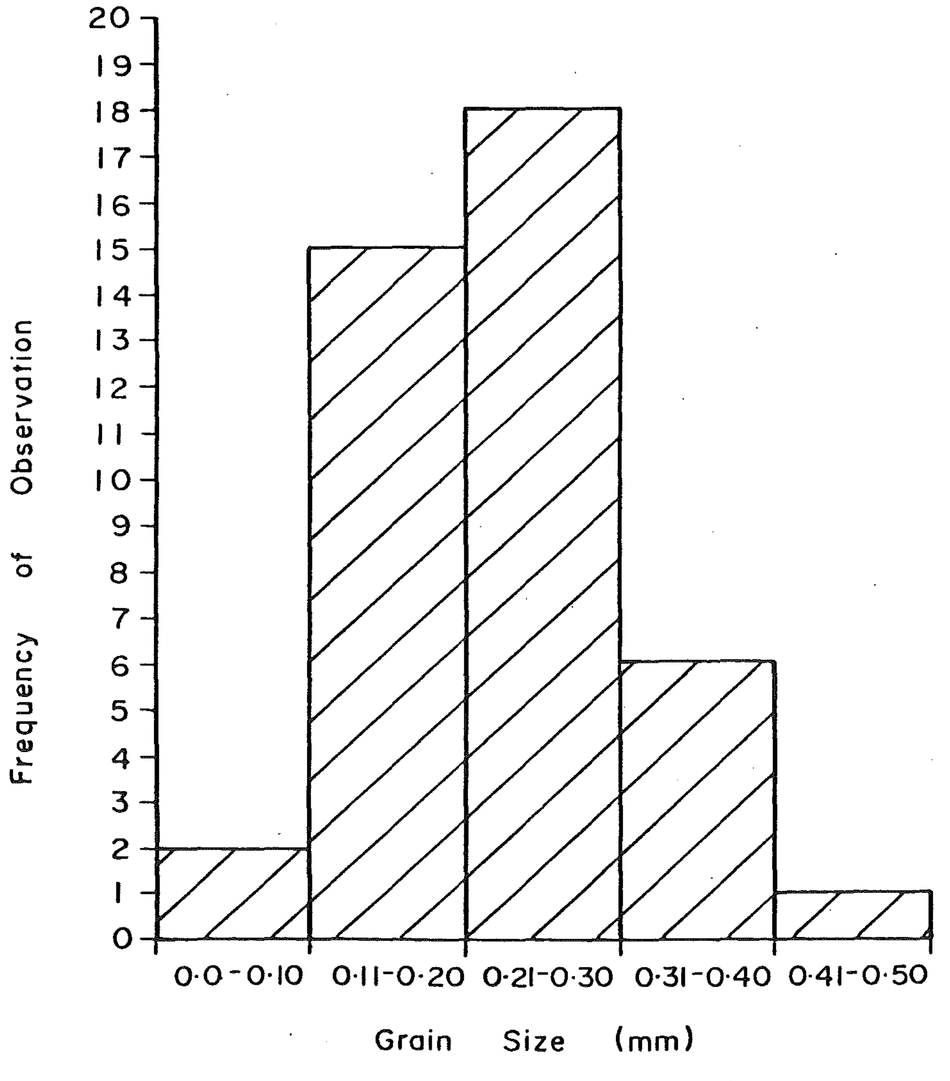


Figure 4.2 GRAIN SIZE OF WIND DEPOSITED SNOW. Mean grain diameter of the forty-two snow strata described in figure 4.1, mean = 0.24 mm, approximately one order of magnitude less than original crystal forms.

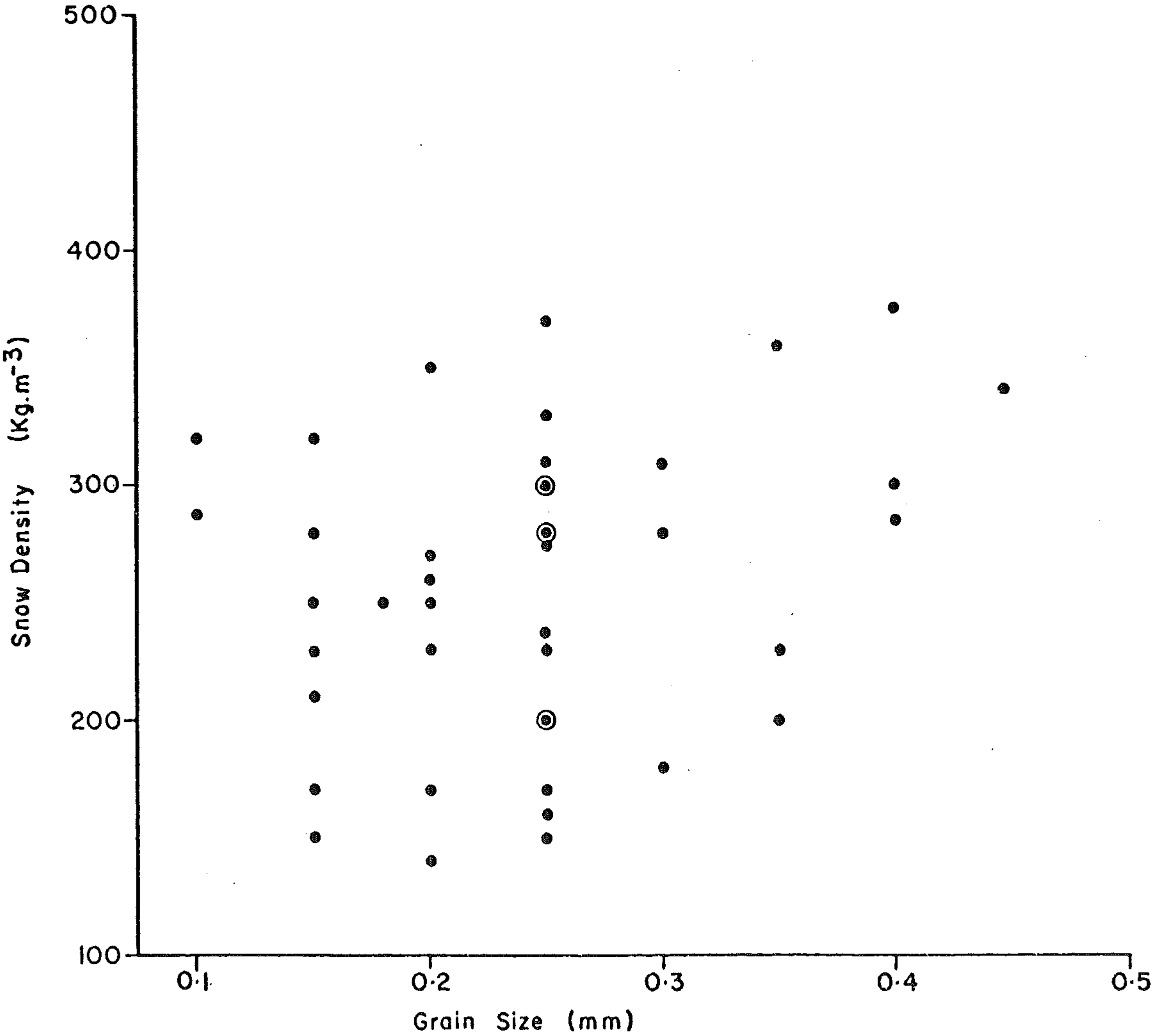


Figure 4.3 SNOW DENSITY VERSUS GRAIN SIZE OF WIND DEPOSITED SNOW. Circles indicate multiple observations and the column distribution of the points is due to a 0.05 mm classification of grain size.

wind packed snow is also known to have relatively high mechanical strength. A relationship of the form

$$\text{Log } R = a + b\,\rho_s \tag{4.1}$$

where R is ram resistance (kg); a is the regression intercept; b is the regression coefficient; ρ_s is snow density (kg m^{-3}),

has been established for polar snow by Bull (1956) and by a variety of researchers for alpine snow [Keeler and Weeks (1968); Martinelli (1971); Weir and Owens (1981)]. A similar form of the equation was derived from the 1979-1980 snow pit record for the Craigieburn Range based on 107 separate samples. Layers of snow noticeably affected by temperature gradient metamorphism (discussed in the following section) or which had a high free water content were not included. This large sample was broken into two smaller samples, one containing only those layers described above as wind deposited and the other, all other layers found at depth within the snowpack. If wind packing leads to greater strength for given densities than normal metamorphic-compaction processes, this should be revealed in different coefficients for equation 4.1. A summary of the regression and correlation coefficients for equation 4.1 appears in table 4.1.

The results from this study compare quite favourably with the values derived by other researchers. However, the equation for the wind packed sample does not appear to reflect greater ram numbers (mechanical strength) than snow of similar density from other areas. Further to this point,

INVESTIGATOR	a	b	r	SNOW CONDITIONS & DENSITY (kg m^{-3})	LOCATION
BULL (1956)	-0.611	.0054	0.80	ALL; 100-510	GREENLAND
KEELER AND WEEKS (1967)	-0.845	.0064	0.94	ALL; 150-430	MONTANA
MARTINELLI (1971)	-0.428	.0054	0.85	ALL; 40-450	COLORADO
MARTINELLI (1971)	-0.463	.0042	0.68	ALL; 100-430	COLORADO
WEIR AND OWENS (1981)	-0.477	.0050	0.85	ALL; 150-460	MT HUTT
THIS STUDY	-0.408	.0046	0.79	ALL; 100-510	CRAIGIEBURN RANGE
THIS STUDY	-0.485	.0051	0.83	SURFACE WIND DEPOSITS; 140-375	CRAIGIEBURN RANGE
THIS STUDY	-0.392	.0045	0.78	INTERIOR SNOW 100-510	CRAIGIEBURN RANGE

Table 4.1 REGRESSION EQUATIONS RELATING SNOW DENSITY TO MECHANICAL STRENGTH. Mechanical strength of the snowpack was represented by the mean Ram number for a given strata, similarly, density values were also an average. The regression and correlation coefficients are as defined in equation 4.1.

table 4.2 illustrates the ram values derived for specific density values. The wind affected layers seem to be characterized by mechanical strength slightly higher than those for the interior of the snowpack. However, they are in most cases not significantly different in strength than the layers reported by other researchers. A t-test was used to test for the significance (95% confidence level) of the difference between the regression coefficients derived in this study and revealed that there was no significant difference between the wind-deposited and interior snow samples.

	SNOW DENSITY (kg m^{-3})			
	100	200	300	400
BULL (1956)	0.9	3.0	10.2	35.5
KEELER AND WEEKS (1967)	N.D.	2.7	11.9	51.9
MARTINELLI (1971)	1.3	4.5	15.6	54.0
MARTINELLI (1971)	0.9	2.4	6.3	16.5
OWENS AND WEIR (1980)	1.1	3.3	10.5	33.3
THIS STUDY:				
ALL DENSITIES	1.1	3.3	9.4	27.0
WIND PACKED SNOW	N.D.	3.4	11.1	35.9
INTERIOR SNOW	1.1	3.2	9.1	25.6

Table 4.2 MECHANICAL STRENGTH FOR VARIOUS SNOW DENSITIES. Mechanical strengths are in ram numbers (kg) based on the equations listed in table 4.1. N.D. refers to 'not defined' by the respective range of densities used in the original regression analysis.

In summary, the strength values for given densities of wind deposited snow does not significantly differ from snow of similar densities found throughout the remainder of the snowpack. Wind packing effectively produces a surface snow similar in density, grain size and strength to that which forms from metamorphic processes, especially that of equi-temperature metamorphism. The different types of metamorphism are discussed next.

4.4 METAMORPHIC PROCESSES BY VAPOUR TRANSFER

4.4.1 Background

Once snow has been deposited on the ground, either in calm or windy conditions, structural changes occur within the ice matrix. A majority of these changes are due to vapour transport and have been classified by Eugster (1952) into two principal types, constructive and destructive metamorphism. The distinction between the two primarily focuses on whether a critical temperature gradient and hence vapour pressure gradient exists within the snowpack. More recently the terms temperature gradient (TG) and equi-temperature (ET) metamorphism (referring respectively to constructive and destructive metamorphism) have come into common use (Sommerfield and La Chapelle 1970).

Changes related to ET metamorphism were first documented by Bader et al. (1939), who postulated that if snow is considered in a thermodynamic sense, ET metamorphism could be viewed as attempting to minimize free energy. Some of the highest states of free energy exist in new snow crystal forms characterized by high surface area to volume ratios. The first stage of ET metamorphism involves sublimation of water (Hobbs and Radke 1967) such that sharp corners and edges on these crystals become rounded and a general dissection of the original crystal shape occurs. This process eventually produces much smaller grain sizes down to a minimum of approximately 0.1 mm and is accompanied by a process of sintering in which bonds grow at contact points between grains (Ramseier and Keeler 1966). Hence, in many aspects the overall structure of snow formed from the first stage of ET metamorphism is very similar to that developed by wind compaction.

As ET metamorphism continues, larger grains develop and continue to grow at the expense of smaller ones. With this gradual increase in average grain size, the overall rate of change decreases until an upper limit of 1 mm is reached (Sommerfield and La Chapelle 1970). Hobbs and Radke (1967) point out that in the latter stages of ET metamorphism, when density values approach 400 kg m^{-3}, volume diffusion through the grains replaces the movement of water in the vapour phase as the dominant transport mechanism. In a recent review, Male (1980) showed that vapour transport mechanisms are strongly influenced by grain surface curvature, surface stress and crystal structure.

As the name equi-temperature metamorphism implies, it occurs in snowpacks which do not have an appreciable temperature gradient. However, as the vapour pressure over ice is known to be higher at temperatures nearer freezing, the rates of ET metamorphism are maximized in warm snowpacks and follow a log time linear densification at colder temperatures [Gow (1975); Hobbs (1965, 1968)]. For most purposes, ET metamorphism can be seen to almost cease at -40^{o}C (La Chapelle 1969). Hence, for very cold polar and alpine snowpacks, the time required for natural settling and densification produced by ET metamorphism is highly protracted.

In contrast to equi-temperature metamorphism, temperature gradient metamorphism is known to occur in snow where appreciable vertical temperature differences prevail. Strong snow temperature gradients exist primarily because of large differences between the ground and air temperatures. Since, in most climates, ground temperatures are at or reasonably near the freezing point, air temperatures primarily control the snow temperature gradient. If the gradient persists long enough, vapour transfer across air spaces and through grains (La Chapelle and Armstrong 1977) will cause the recrystallization of new crystal forms. Depth hoar, the end result of TG metamorphism has been referred to as the coarsest grained snow structure which may form in the absence of liquid water. The distinctive shape of depth hoar crystals is probably why they were recognized relatively early. Seligman (1936) records their first appearance in polar regions over 120 years ago and more recently Paulcke (1932) noted depth hoar in alpine areas.

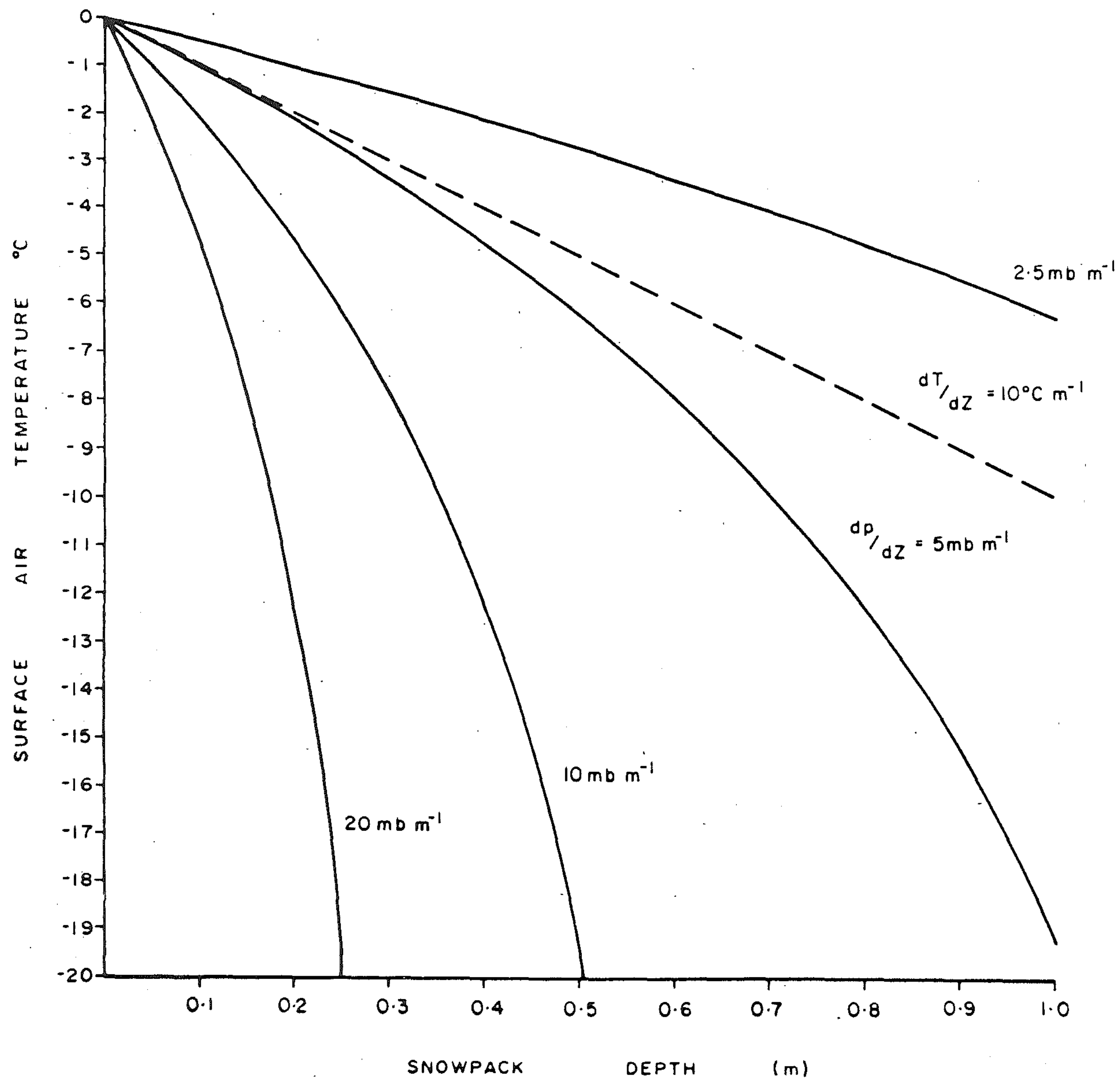
Depth hoar has been considered to be a phenomenon generally limited to the basal layers of cold continental, polar and high alpine snowpacks (La Chapelle and Armstrong 1977). So regular is the formation of basal depth hoar that many researchers have used its presence to separate annual strata in snowcover (Benson 1962). However, Benson (1979) and Trabant and Benson (1972) working in the interior of Alaska report that in extreme cases entire snowpacks can be predominantly depth hoar crystals. It is also known to occur above and below ice layers within a snowpack (Perla and Martinelli 1976), although the reasons for these particular locations are still unclear.

The initiation of temperature gradient metamorphism is most common in snow with densities between 100 and 250 kg m^{-3}. As reported by Giddings and La Chapelle (1962), densities in excess of 300 kg m^{-3} seem to restrict the vapour transfer process and hence TG metamorphism has difficulty functioning. The initial formation rates have also been observed to be more rapid than subsequent growth rates. In addition, as for ET metamorphism, since vapour pressure over ice is exponentially related to temperature, vapour transfer and hence growth rates increase with temperature (de Quervain 1973). Giddings and La Chapelle (1962) have calculated that the vapour flux is reduced by a factor of 2 for every $8^{o}C$ decrease in the snow temperature. Since the warmest temperatures are normally found at the base of a snowpack, this is the location where depth hoar growth rates are usually the most rapid.

Both ET and TG metamorphism can occur in the same snowpack at the same time depending on the variation of the temperature gradient within the snow column. Accumulated experience has pointed to a critical temperature gradient of $10^{o}C\ m^{-1}$ which determines whether TG or ET metamorphism will occur. However, La Chapelle and Armstrong (1977) have indicated that, as a result of the non-linear relationship between temperature and vapour pressure, temperature gradients are not valid indices for separating TG and ET metamorphism. Instead, they define a critical pressure value of 5 mb m^{-1} to be used in separating the two processes. This author has constructed figure 4.4 to illustrate the relationship between temperature and vapour pressure gradients over a variety of air temperatures (assuming a ground temperature at $0^{o}C$) and snowpack depths. Differences between the two gradients can be demonstrated in figure 4.4 by considering a 1 m snowpack with a $10^{o}C$ temperature difference existing between the surface snow and ground temperatures. This would produce the reported temperature gradient of $10^{o}C\ m^{-1}$ but would be considerably less than the 5 mb m^{-1} vapour pressure gradient required for TG metamorphism to occur. A temperature gradient of over $19^{o}C\ m^{-1}$ would in fact be necessary. The reason a temperature gradient of $10^{o}C\ m^{-1}$ has received general acceptance is that most depth hoar forms at the base of snowpacks where snow temperatures are near the freezing mark, the point at which the difference between the two gradients is minimized.

The end result of the recrystallization process from TG metamorphism is an increase in grain size (frequently to 5 mm) and in many cases a decrease in snow density. At full development, depth hoar in the United States Rocky Mountains and the Swiss Alps has been found to be 300 kg m^{-3} (Giddings and La Chapelle 1962) and a similar value has been reported for the Greenland ice sheet (Benson 1962). However, in the interior of Alaska, Benson (1979) notes mean values of 200 kg m^{-3}, which he claims to be some of the lowest anywhere.

Figure 4.4 RELATIONSHIP OF TEMPERATURE AND VAPOUR PRESSURE GRADIENTS. The 5 mb m^{-1} line is the critical value for separating between ET and TG metamorphism. The $10^{o}C\ m^{-1}$ line offers a reasonable approximation to approximately $-5^{o}C$ but is totally inaccurate at colder temperatures.


SURFACE AIR TEMPERATURE °C
0
-1
-2
-3
-4
-5
-6
-7
-8
-9
-10
-11
-12
-13
-14
-15
-16
-17
-18
-19
-20
0·1
0·2
0·3
0·4
0·5
0·6
0·7
0·8
0·9
1·0
SNOWPACK DEPTH (m)
2·5 mb m⁻¹
dT/dZ = 10°C m⁻¹
dp/dZ = 5 mb m⁻¹
10 mb m⁻¹
20 mb m⁻¹

One of the important structural differences between snow crystals-grains formed by the two types of metamorphism is that TG metamorphism is not accompanied by a sintering process. This lack of inter-granular bonding greatly increases the viscosity of snow and is frequently identified as the key factor in the production of climax avalanches (Perla and Martinelli 1976).

4.4.2 Observations in the Craigieburn Range

Two approaches were used to assess the nature of metamorphic processes in the snowpack of the Craigieburn Range. The first involved indirect meteorological evidence and the second direct observation of snowpack structure.

4.4.2.1 Long Term Snowpack Temperature Gradients

The presence of either ET or TG metamorphism within a snowpack depends primarily on the difference between the surface snow and ground temperatures. As will be demonstrated in chapter six, the ground temperatures at the SB site are usually at or near the freezing point. Hence, as established by La Chapelle and Armstrong (1977), the long term effect of mean daily air temperatures determines the temperature gradient within a snowcover as a whole. Akitaya (1974) suggested that the mean monthly snowcover temperature gradient could be calculated by:

$$\frac{d\bar{T}_s}{dz} = \frac{\bar{T}_a}{\bar{d}_s} \qquad (4.2)$$

where $\frac{d\bar{T}_s}{dz}$ is the mean monthly snowcover temperature gradient ($^{o}C\ m^{-1}$); $\bar{T}_a$ is the mean monthly air temperature (^{o}C); $\bar{d}_s$ is the mean monthly snow depth (m).

Equation 4.2 was used to calculate the long term mean monthly temperature gradients for the SB site over an eight year period from 1973 to 1980. Mean monthly temperatures were calculated from monthly averages of daily maximum and minimum air temperatures and mean monthly snow depths were the averages of daily values within each month. The results of the analysis appear in table 4.3.

	MEAN MONTHLY SNOWPACK TEMPERATURE GRADIENT ($^{o}C\ m^{-1}$)			
YEAR	JUNE	JULY	AUGUST	SEPTEMBER
1973	2.1	3.5	1.2	+
1974	1.8	3.3	3.2	+
1975	*	1.5	1.0	+
1976	5.7	1.2	0.8	0.6
1977	3.4	1.5	1.4	N.D.
1978	7.8	1.0	+	+
1979	+	N.D.	1.7	+
1980	2.8	N.D.	1.0	+
AVERAGE	3.9	2.0	1.5	N.D.

Table 4.3 CALCULATED MEAN MONTHLY SNOWPACK TEMPERATURE GRADIENTS. Values refer to the 1550 m SB climate station and are derived from mean air temperatures and snow depths according to equation 4.2. '+' denotes a positive temperature gradient; 'N.D.' denotes insufficient data; '*' denotes snowpack not established by the respective date.

Generally, mean monthly snow temperature gradients averaged over the eight years remained below 4.0°C m^{-1} with a decrease of 50% from June to August. Similarly, for most individual years, the maximum gradient occurred early in the season and was minimized by September or reversed because of positive mean monthly air temperatures. Only during two periods, both in June, did the temperature gradient exceed 5.0°C m^{-1} (snow depth records for these years appear later in the chapter and in section 6.6.2.). A majority of the results fall into the low range of temperature gradients suggested by Yosida (1963) to be common for most snowcovers and all are significantly less than the 'very high' gradient of 18°C m^{-1} reported by de Quervain (1958) for alpine areas.

Since all values for $\frac{dT_s}{dz}$ were less than the critical temperature gradient of 10°C m^{-1}, this would suggest equi-temperature metamorphism would dominate the snowpack structure at the SB site. However, these results are site specific and do not consider spatial variability in snowcover depth and variations in air temperature, specifically with elevation.

4.4.2.2 Influence of Elevation

Mean snowpack temperature gradients would be expected to increase with elevation whenever the increase in snow depth is not sufficient to offset the decrease in temperature with elevation. Figure 4.5 depicts a nomogram illustrating the dependence of temperature gradients on various air

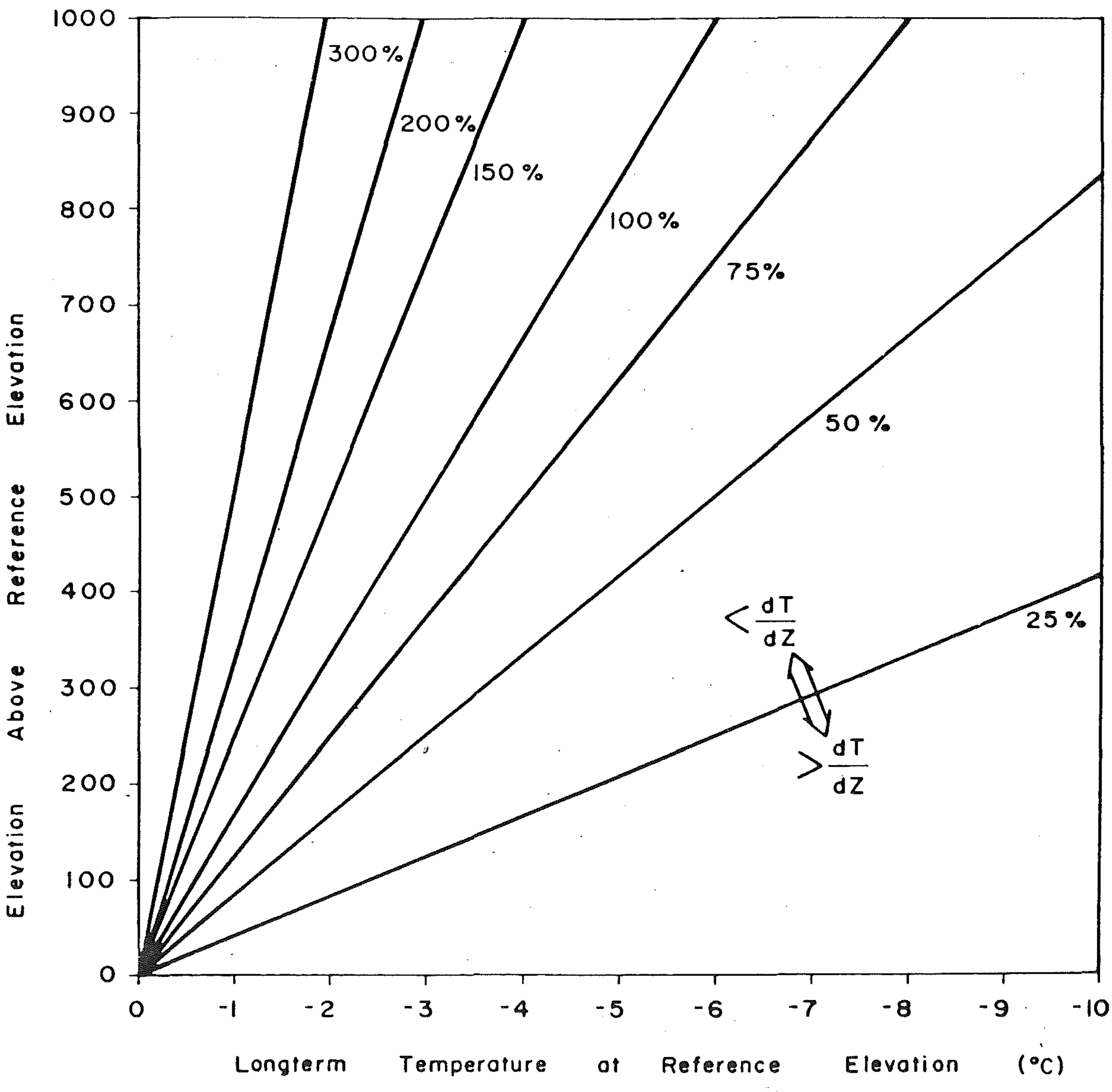


Figure 4.5 CHANGES IN SNOWPACK TEMPERATURE GRADIENTS WITH ELEVATION AND SNOW DEPTH. Percentage lines are based on equation 4.3 and a full explanation of the use of the nomogram is made in the text.

temperatures and snow depth increases with elevation. Temperature changes are based on a lapse rate of $6^{o}C\ km^{-1}$, suggested by Obled and Harder (1979) to be representative of long term conditions in alpine areas. The percentage lines indicate the increase in snow depth from a reference site to a higher elevation zone which would produce the same snowpack temperature gradient as at the reference site. The percentage ($\%\ d_s$) for any given combination is based on

$$\%\ d_s \quad = \quad (E_b - E_a)\ \ell/T_B \qquad (4.3)$$

where E_b is the elevation of a reference site (m); E_a is the elevation of an upper elevation site (m); T_B is the long term temperature at E_B (^{o}C); ℓ is the long term mean lapse rate ($6^{o}C\ km^{-1}$).

As an example, assuming the long term winter temperature at the SB site (1550 m) is approximately $3^{o}C$ below freezing, a 49% increase in snow depth would be required in the upper accumulation zones at 1800 m to produce the same overall snowpack temperature gradient as at the SB site. A greater percentage increase in snow depth between the two elevations would decrease the gradient and a smaller increase would increase the gradient. Although no accurate elevation-snow depth information exists for the study area, increases of over 100% between 1500 and 1800 m have been frequently observed. In one particular north-westerly storm (figure 4.6), the increase was measured at almost 300% over this elevation range. A number of

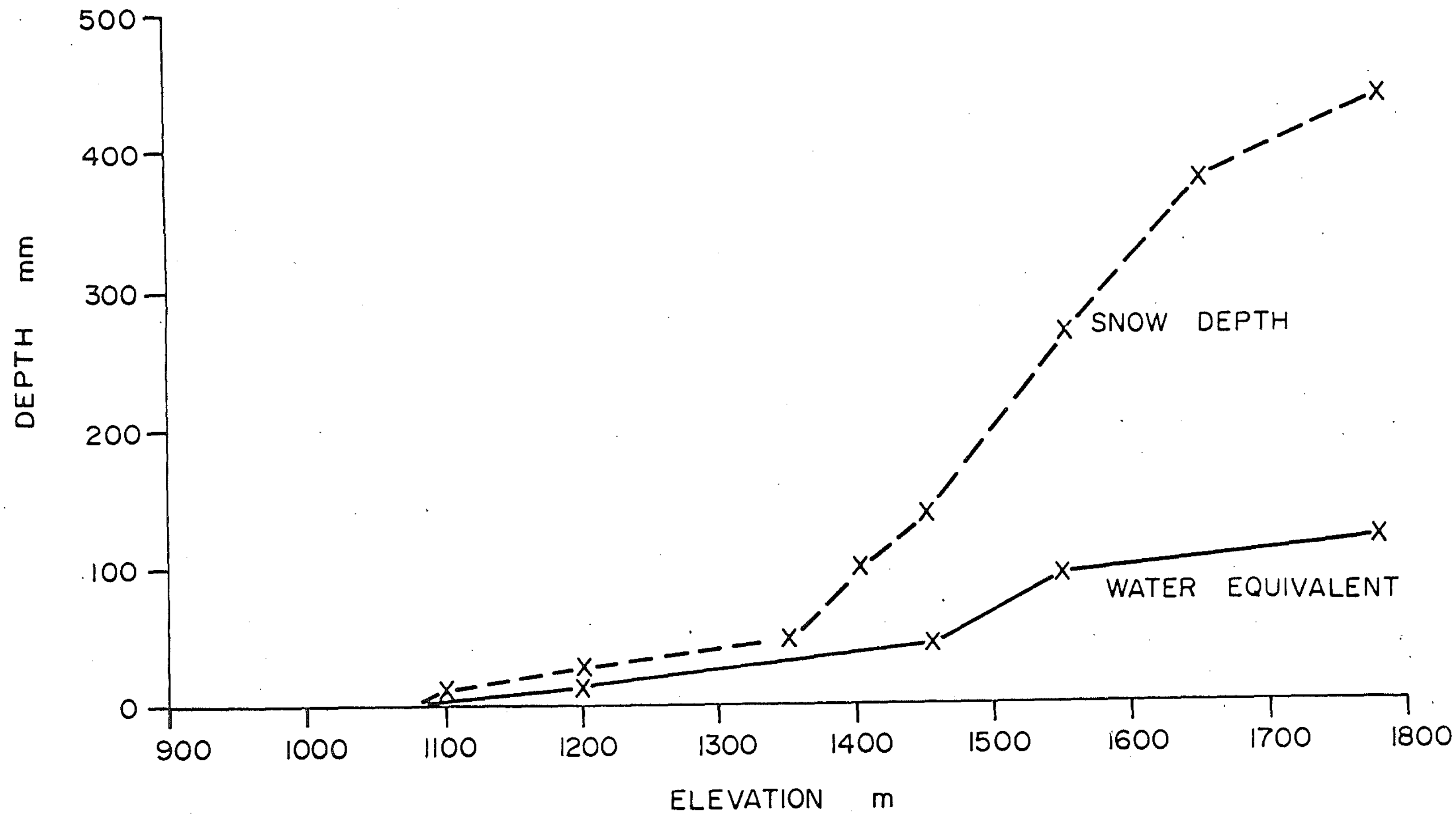


Figure 4.6 SNOW DEPTH INCREASE WITH ELEVATION AFTER A NORTH-WESTERLY STORM. Considerable variation existed in the density composition at various elevations. The mean density at 1200 m was 450 kg m^{-3} at 1450 m; 320 kg m^{-3} at 1550 m and, 263 kg m^{-3} at 1780 m.

articles have been published which deal with snow depth-elevation relationships in New Zealand [for example Archer (1970); Chinn (1969); Fitzharris (1972, 1976, 1977); Morris and O'Loughlin (1965); O'Loughlin (1969a); Weir (1979)] but the results are very site and time specific. As noted by Fitzharris (1977), a large amount of the variation in snow accumulation with elevation is controlled by the number and freezing levels of storms, melt and orographic increase of precipitation. However, small percentage increases in snow depth are known to occur in areas where wind deflation plays a major role in snow accumulation. Weir (1979) working on Mt Hutt, in fact measured a decrease in snow depth at the higher elevations because of large scale wind erosion. In such deflation zones, the temperature gradient would be steepened. An example of one such location is described later in this chapter.

4.4.2.3 Daily Snowpack Temperature Gradients

The previous discussion has considered mean snowpack temperature gradients over the long term but considerable variation is expected to exist in these gradients over shorter time frames. Cold periods lasting much less than a month could quite conceivably establish a temperature gradient which would allow TG metamorphism to occur. La Chapelle and Armstrong (1977) have considered the length of time required to form TG crystals under various vapour pressure gradients. Based on their results, the time required for recyrstallization can be described by:

$$t = \psi / (\frac{dP_v}{dz}) \qquad (4.4)$$

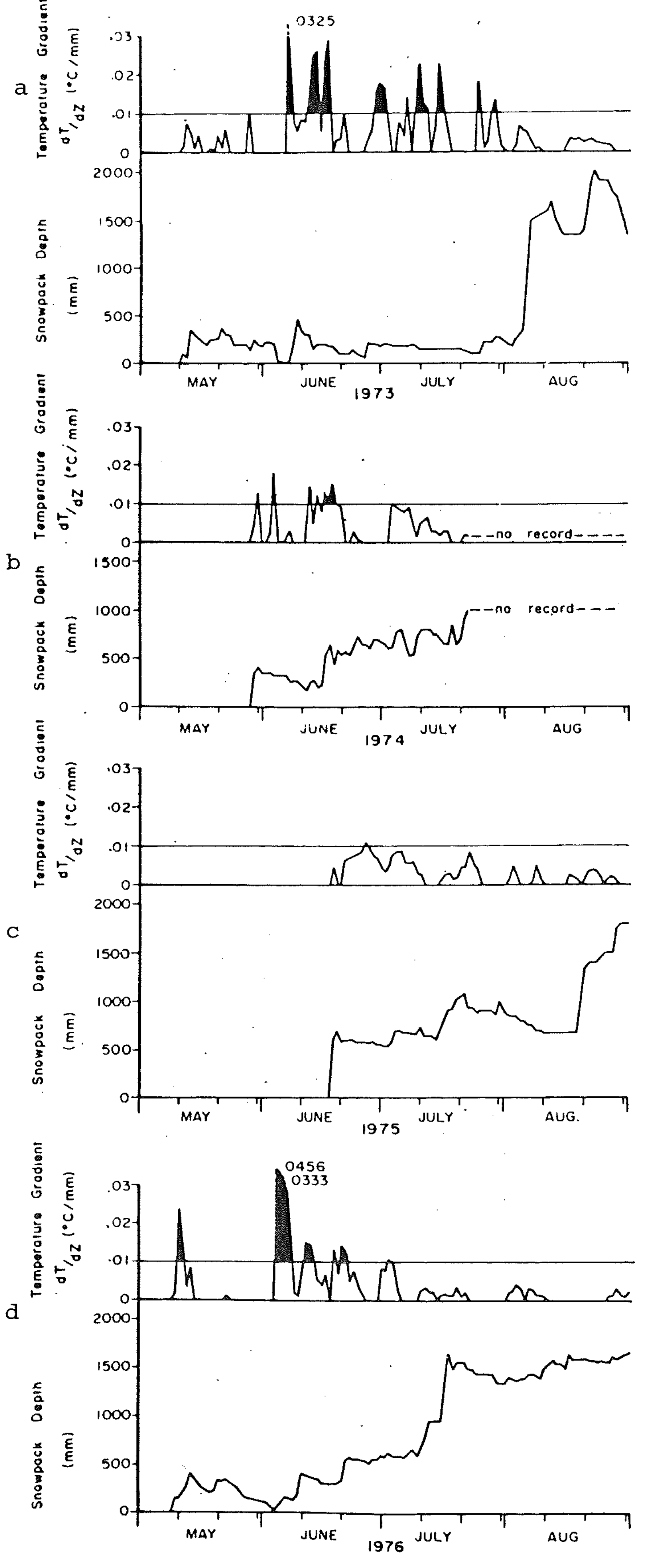
where t is the time in days; ψ is the coefficient of recrystallization; $\frac{dP_v}{dz}$ is the vapour pressure gradient (mb mm^{-1}).

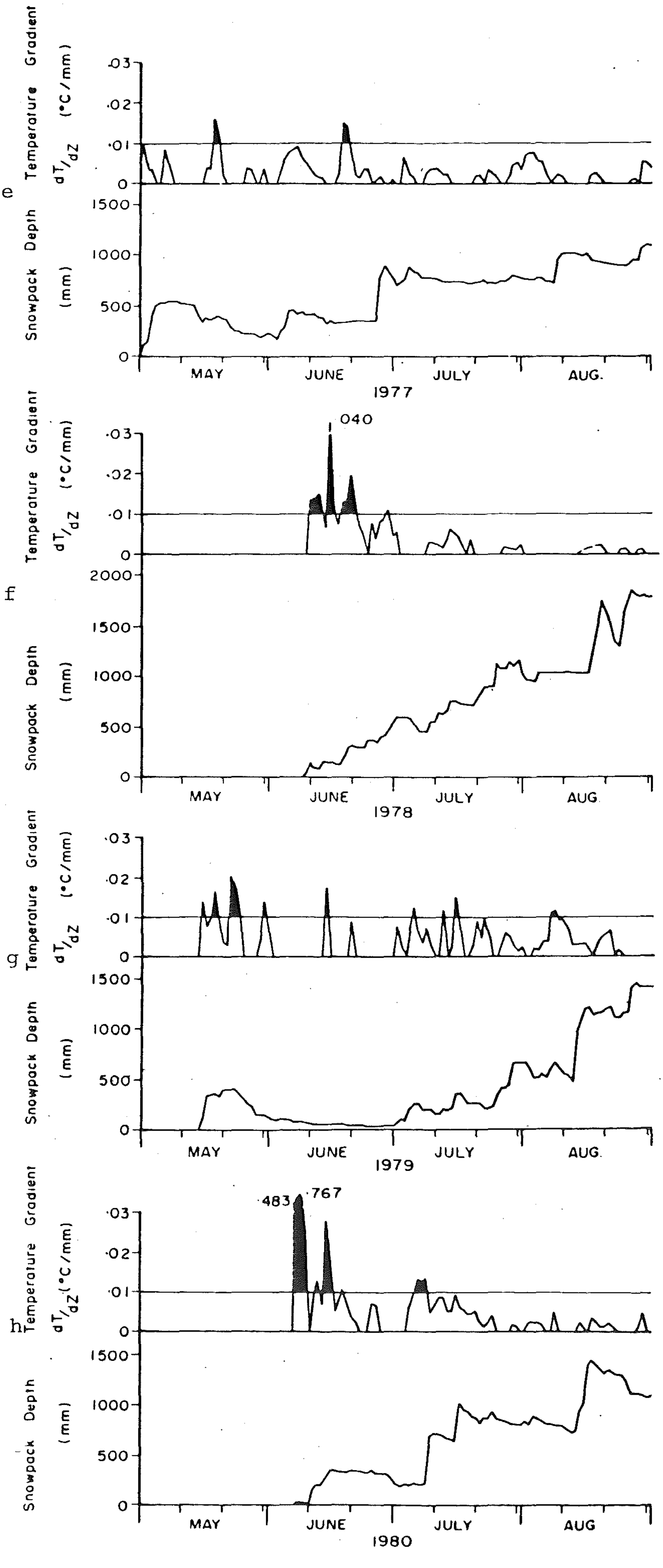
A mean value of ψ for early TG metamorphism is 0.05 (La Chapelle and Armstrong 1977). Based on this value and the minimum critical vapour pressure gradient of 5 mb m^{-1} discussed in section 4.4.1, ten days would be required to produce early stages of TG crystals. Equation 4.4 was derived from observations of TG metamorphism in light density snow (50 - 150 kg m^{-3}) and hence, longer development times would be expected to be required in higher density snow because of the greater restriction to the flow of vapour.

Since metamorphic processes may occur on a time scale which would not be revealed in the mean monthly temperature gradients, daily gradients based on snow depth and mean air temperature were also calculated. The results for the years 1973 to 1980 appear in figures 4.7a-h. Only the winter periods up to the end of August are included because the warmer temperatures and large snow depths in the latter part of the season eliminated the potential for large snowcover temperature gradients.

In all years, there were no periods in which a strong temperature gradient of greater than 10°C m^{-1} persisted longer than four or five days. Only when the snow depth

Figure 4.7a-h CALCULATED DAILY SNOWPACK TEMPERATURE GRADIENTS AND SNOW DEPTH 1973-1980 (2 PAGES). Black areas indicate periods when gradient exceeded .01^{o}C mm^{-1}.


a
Temperature Gradient
dT/dZ (°C/mm)
.0325
.03
.02
.01
0
Snowpack Depth (mm)
2000
1500
1000
500
0
MAY
JUNE
JULY
AUG
1973
b
no record
1500
1000
500
0
MAY
JUNE
JULY
AUG
1974
c
MAY
JUNE
JULY
AUG.
1975
d
0456
0333
MAY
JUNE
JULY
AUG
1976


e
Temperature Gradient
dT/dZ (°C/mm)
.03
.02
.01
0
Snowpack Depth (mm)
1500
1000
500
0
MAY
JUNE
JULY
AUG.
1977
f
.040
.03
.02
.01
0
2000
1500
1000
500
0
MAY
JUNE
JULY
AUG.
1978
g
.03
.02
.01
0
1500
1000
500
0
MAY
JUNE
JULY
AUG.
1979
h
.483
.767
.03
.02
.01
0
1500
1000
500
0
MAY
JUNE
JULY
AUG
1980

was less than 0.6 m did the gradient exceed this value. In some cases, the temperature gradient appears quite large because of very shallow snowcover.

Again, these results would indicate that equi-temperature metamorphism should dominate the snowpack structure at the SB site. However, the observations based on meteorological data consider the snowpack as a whole. In many cases, considerable variation exists in temperature gradients within a snowpack because of differences in inter-snowpack thermal conductivity and the concentration of energy exchange at the snow surface. The next section considers the interior structure of the Craigieburn snowpack based primarily on field observations.

4.4.3 Field Observations in the Craigieburn Range

The following three sub-sections describe the nature, rate and crystal forms of both equi-temperature and temperature gradient metamorphism observed in the snowpack of the Craigieburn Range.

4.4.3.1 ET Metamorphism

As described in section 4.3.2 the surface wind packed snow strata exhibited grain sizes and mechanical strengths similar to those which normally develop from the effects of ET metamorphism. Further, rates of ET metamorphism in these layers were slow in comparison to the rate for the new light density snow deposited under calm conditions. In five cases of snow with an initial density of less than 100 kg m^{-3}, ET metamorphism produced an average increase in density of 47% within a twenty-four hour period; a rate comparable to those reported by Keeler (1967) for other alpine snowpacks. The most rapid increases in the density of new snow occurred during storms in which the air temperatures fluctuated about the freezing mark. In some cases, the density of surface layers 200 - 300 mm in depth increased to a density of over 400 kg m^{-3} within a few hours. These rapid changes were usually assisted by the presence of free water in the snowpack and therefore cannot strictly be considered due to ET metamorphism alone. The role of free water in metamorphism is discussed in section 4.5.

Because of the rapid metamorphism of new snow, it was rare to identify any new snow crystal forms below the immediate surface layers. In contrast, within polar and cold continental snowpacks, rates of ET metamorphism are so slow that the persistence of original crystal forms is often measured in weeks. Although most original crystals within the Craigieburn snowpack disappeared in approximately one to two days, there were two notable exceptions, graupel and surface hoar. Grains of graupel approximately 2 mm in diameter were periodically found in between metamorphosed ET grains only 0.1 to 0.2 mm in diameter. Graupel is known to be largely resistant to ET metamorphism and may persist in a snowcover for extended periods. La Chapelle (1969) has noted graupel remaining in a snowpack for 60 days with little change to the original form. The main reason for this resistance to ET metamorphism is that the size and shape of grapuel grains is the same as which ET metamorphism attempts to produce. However, the main distinction between ET grains and graupel is that the latter form does not possess appreciable inter-granular bonding as which forms from sintering in ET metamorphism. Hence, buried within a snowpack graupel exhibits very low mechanical strength compared to equivalent density ET snow and because of this weakness may act as a sliding layer in avalanching.

The second persistent crystal form, surface hoar, is quite prevalent in the Craigieburn Mountains. During cold clear periods, surface hoar crystals 1 - 5 mm in size, regularly formed and persisted on south facing slopes. These surface crystals frequently accumulated to a thickness of 20 mm or more, but in the higher alpine regions of New Zealand, they have been observed over 150 mm thick (Wills 1980, pers. comm.). Once covered by subsequent snowfalls, surface hoar layers were almost impossible to detect on the wall of a snow pit. As for graupel, buried surface hoar is an effective sliding layer and hence the best way of locating these layers was by a shovel test (appendix A).

Within many climates, the effects of ET metamorphism on specific snow layers may be traced over the entire winter season. Although construction of time profiles were attempted for the Craigieburn snowpack, the complexity of metamorphic processes precluded the accurate tracing of temporal trends for specific layers. In particular, melt-freeze metamorphism frequently blended distinctly different snow strata into single homogeneous snow masses. ET metamorphism rarely produced grains larger than 0.5 mm without the rate of grain growth being hastened by large amounts of free water. For example, in one case a new snow layer was observed to increase in grain size from 0.1 to 1.0 mm by ET metamorphism in eighteen days, but similar increases in grain dimensions were noted to occur in two or three days in the presence of only moderate amounts of free water.

Section 4.5 later in this chapter outlines the major effects of melt-freeze metamorphism on snow structure.

4.4.3.2 TG Metamorphism

In view of the evidence from the climate-snowcover relationships described in 4. 2 it would appear that TG metamorphism should be almost non-existent in the Craigieburn Range. However, during the 1979-80 winters, snow layers noticeably affected by TG metamorphism were consistently found in snow pits spanning an elevation range of 1100 to 1800 m. Figure 4.8 depicts the relative location of such layers within the snowpack. Since a number of the snow pit observations were conducted periodically over the winters in a similar area, they cannot be considered as discrete samples. However, a number of important observations can still be made about the characteristics of the layers. First, most layers affected by TG metamorphism were located near the base of the snowpack, the mean percentage depth being 83. TG layers were observed only five times in the upper 60% of the snowpack. The thickness of snowpacks considered in this analysis ranged from as little as 220 mm to well over 2000 mm, with an overall mean of 1020 mm. The mean thickness of the TG layers was 120 mm, or 12% of the mean snowpack depth.

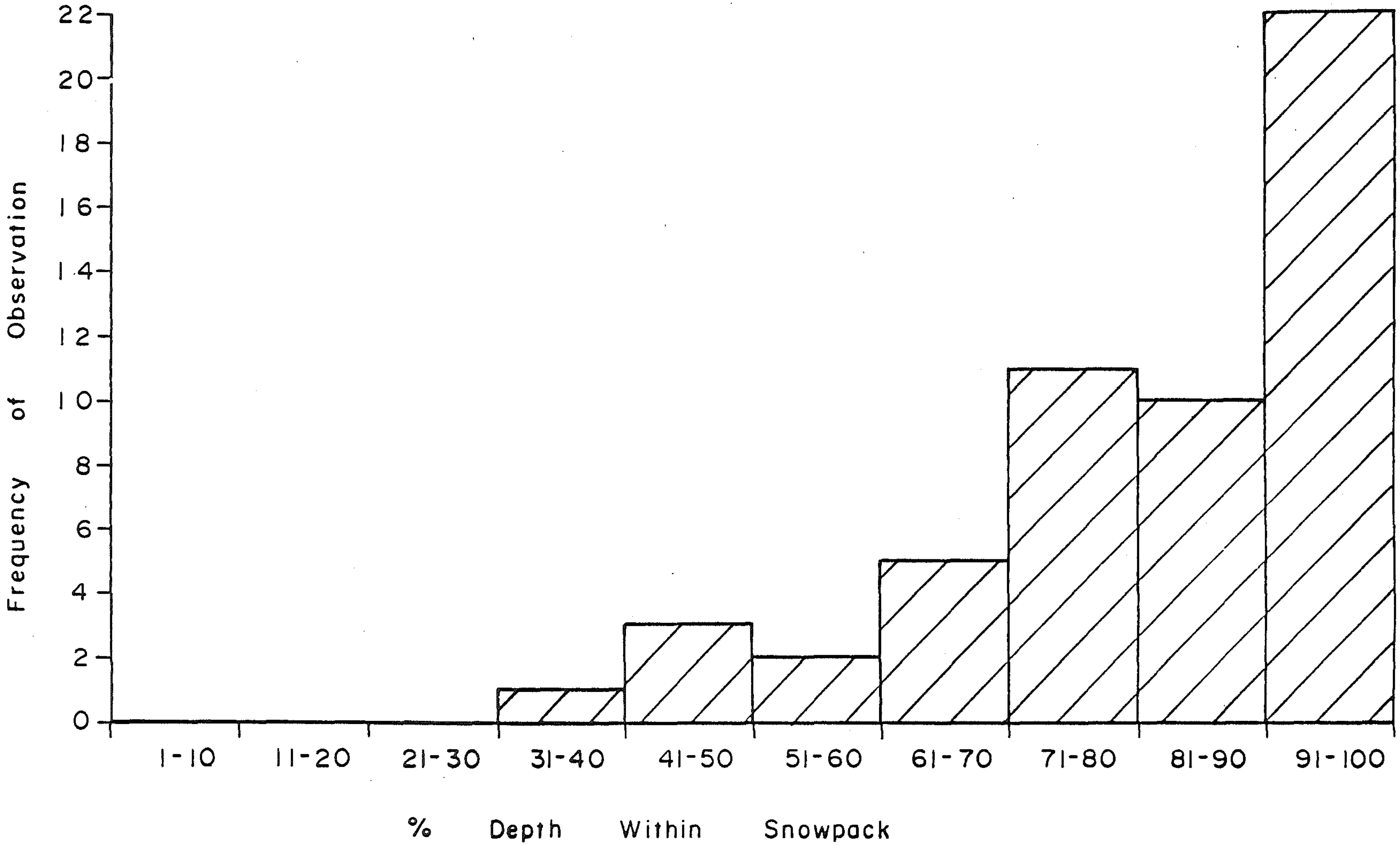


Figure 4.8 DEPTH LOCATION OF LAYERS AFFECTED BY TEMPERATURE GRADIENT METAMORPHISM. The percentage depth refers to the depth of the base of the TG affected layer relative to the total snowpack depth. Only layers which were in active TG formation are included.

Care must be taken in interpreting the above results, especially the depth-location figures. In many cases, TG metamorphism may become established within a snowpack and be subsequently buried by further snowfalls. These additions would tend to obscure the results in figure 4.8 if they were used to identify layers where TG metamorphism is active rather than where TG layers are simply located. However, the layers used in this analysis were ones in which the distinctive sharp edged appearance of TG affected snow was readily apparent. The appearance of such crystals is known to become much more rounded from equi-temperature metamorphism once the critical temperature gradient has been removed, such as which would occur with the addition of new snow. Hence, the process of TG metamorphism must have been active close to the time of sampling and would indicate the results in figure 4.8 are a reasonable approximation of the depth in the snowpack at which TG metamorphism is active.

The densities of the fifty-one layers in which TG metamorphism occurred were also noted (figure 4.9). As pointed out earlier, TG metamorphism has difficulty operating in snow denser than 300 kg m^{-3} (Giddings and La Chapelle 1962) but a full 25% of the observations in this study revealed crystals present in layers having a density above 300 kg m^{-3}. A majority of the layers in which TG metamorphism occurred was between 250 and 300 kg m^{-3} which may simply reflect the overall mean density of the Craigieburn snowpack. This aspect is discussed further, in a later section of this chapter.

The above densities refer only to layers where TG

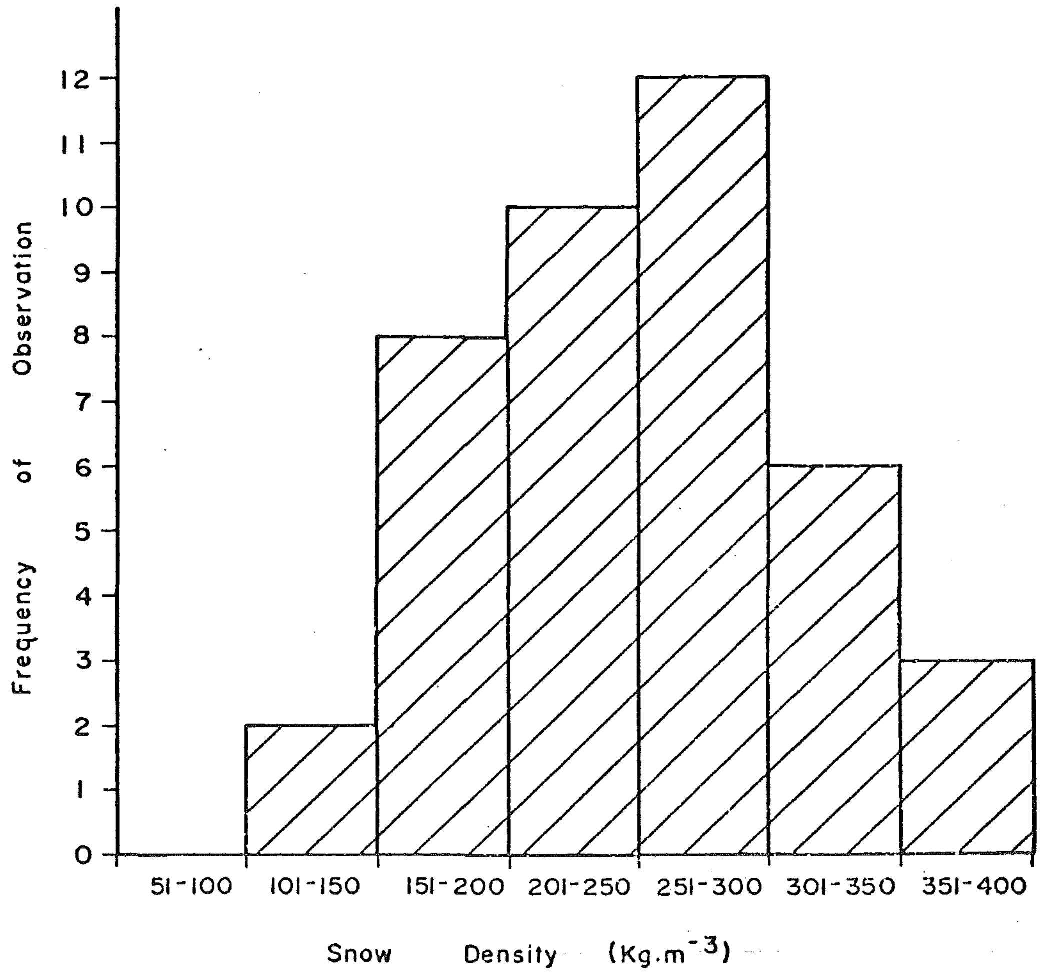


Figure 4.9 DENSITY OF LAYERS AFFECTED BY TEMPERATURE GRADIENT METAMORPHISM. The sample set is the same as in figure 4.8.

metamorphism was present but another important consideration is the density of layers where TG metamorphism was the dominant metamorphic process. Such layers were classified as those where fully developed depth hoar crystals were present and where the mean crystal size exceeded 1.0 mm. Only ten layers fulfilled this criteria ranging in density from 180 to 330 kg m^{-3} and with a mean of 281 kg m^{-3} (standard deviation = 54). The mean density is very similar to the fully developed depth hoar densities previously mentioned for the Rocky Mountains, Switzerland and Greenland. More important, however, is the minimum density of 180 kg m^{-3} which is only 30 kg m^{-3} more than the minimum density of TG crystals reported by Benson (1979) for the Arctic tundra and 20 kg m^{-3} less than the mean density of depth hoar for the interior of Alaska, which he describes as the lowest anywhere. Fortunately a well documented temperature gradient record existed for the location at which this low density depth hoar was discovered.

The snow pit record in figure 4.10 illustrates the snow stratigraphy in which the depth hoar was observed. The location was at a near ridge position at an elevation of 1650 m directly beside the number six thermistor station on the Mt Cockayne temperature profile. The layer was approximately 150 mm thick, located at the very base of a 0.72 m thick snowpack, and was comprised of well developed depth hoar crystals ranging from 1.0 to 3.0 mm in diameter. The upper 200 mm of the snowpack had been deposited by snowfalls a few days prior to August 10, 1979, the day of the snow pit profile. The temperature gradient through the

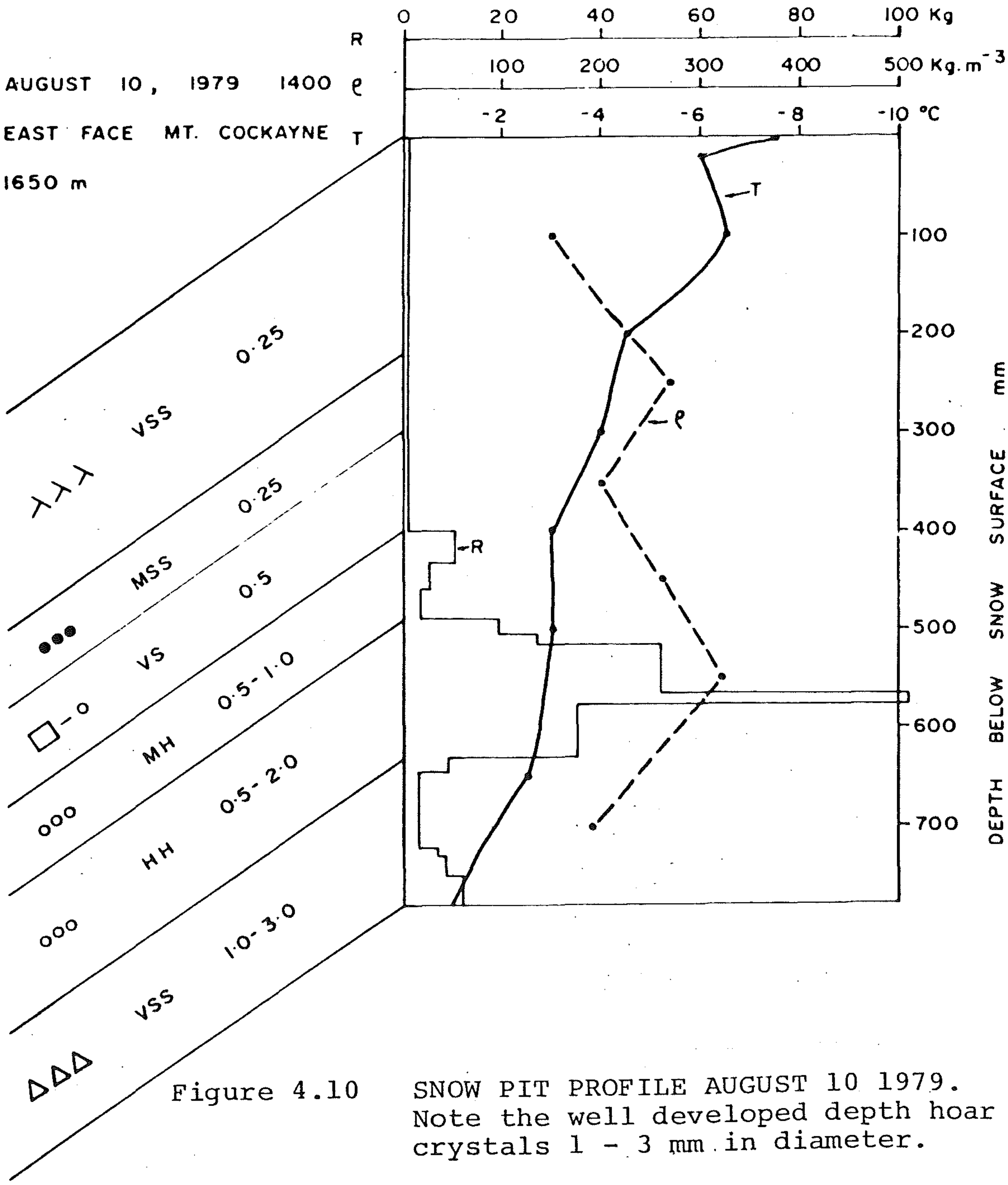
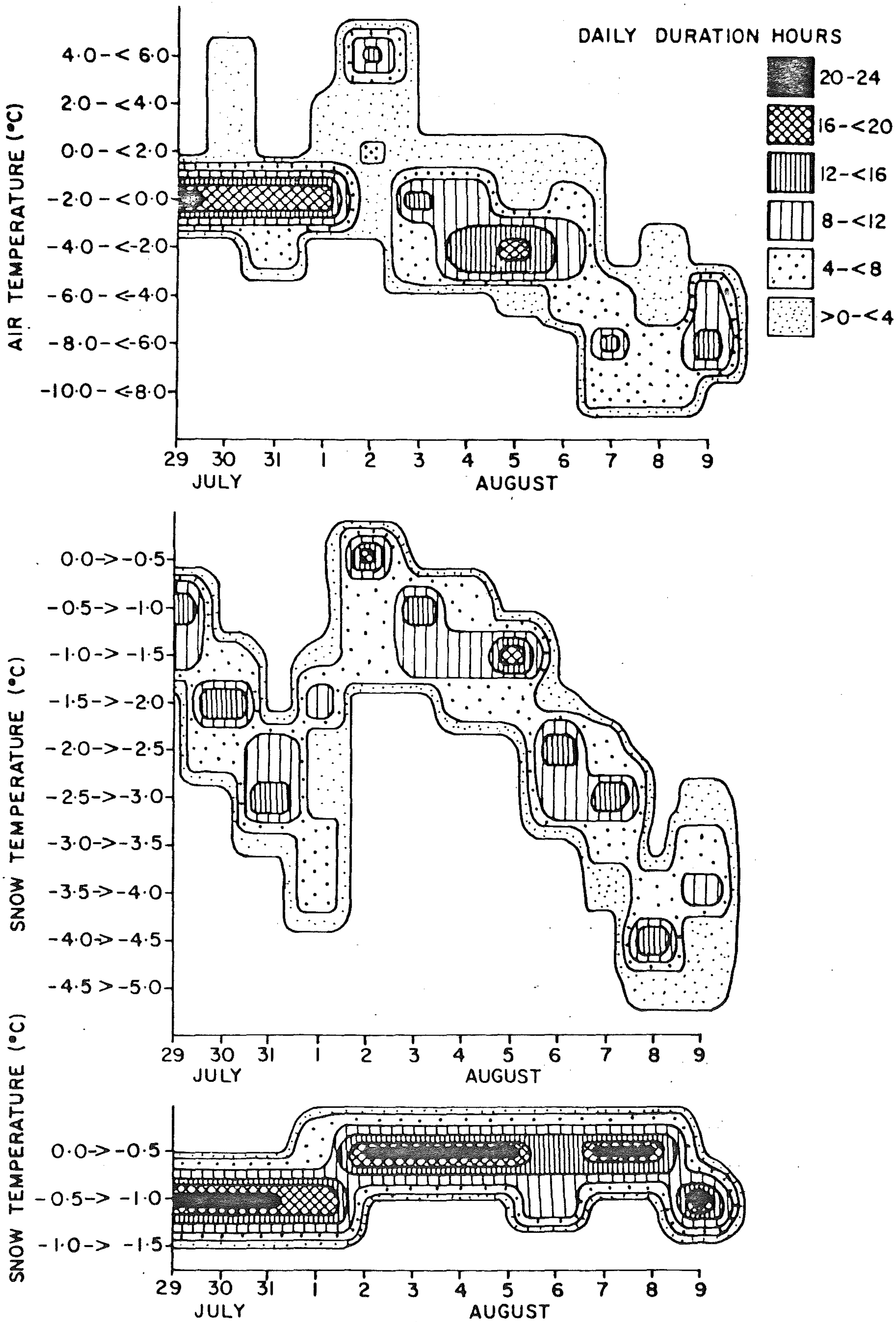


Figure 4.10 SNOW PIT PROFILE AUGUST 10 1979. Note the well developed depth hoar crystals 1 - 3 mm in diameter.

depth hoar layer on August 10 was $10^{o}C\ m^{-1}$, corresponding to a vapour pressure gradient of 4.4 mb m^{-1} just below the critical value of 5 mb m^{-1}. However, the major period of TG metamorphism would have occurred prior to this date. A temperature record from nearby number six thermistor station was available for this earlier period. Both the above ground (+100 m, +300 mm) thermistors at the 1600 m site became snow covered at the end of July and subsequently provided a record of snow temperatures. This record, in addition to that for the +1500 mm thermistor, which indicated air temperatures, are presented in figure 4.11. As illustrated, the trend of temperature in the upper snow thermistor (+300 mm) reflects that of the air thermistor (+1500 mm) while the lower sensor (+50 mm) remains relatively constant near the freezing mark. By subtracting the temperatures between the two snow thermistors, an approximate temperature gradient through a 250 mm section of the lower part of the snowpack can be obtained. Assuming saturation vapour pressure over ice, these may temperatures may be converted to vapour pressure gradients.

Although the snowpack was relatively isothermal on July 29 to July 31, an appreciable gradient in excess of the critical 5 mb m^{-1} had developed. However, by August 2, a near isothermal state was again present and appreciable changes did not occur until August 5 with the decrease in air temperatures. A strong gradient, ranging up to 7.5 mb m^{-1} ($18^{o}C\ m^{-1}$) then persisted until the time of the snow pit observations. Hence, although some depth hoar development was possible early in the period (July 31 - August 1), the

Figure 4.11 SNOW AND AIR TEMPERATURES DURING DEPTH HOAR FORMATION. Daily duration values (hr) refer to the time each sensor was within a particular temperature range. The upper diagram is for the +1500 m thermistor and the two lower ones refer to the +300 and +50 mm thermistors covered by snow.


DAILY DURATION HOURS
20-24
16-<20
12-<16
8-<12
4-<8
>0-<4
AIR TEMPERATURE (°C)
4·0-<6·0
2·0-<4·0
0·0-<2·0
-2·0-<0·0
-4·0-<-2·0
-6·0-<-4·0
-8·0-<-6·0
-10·0-<-8·0
29 30 31 1 2 3 4 5 6 7 8 9
JULY
AUGUST
SNOW TEMPERATURE (°C)
0·0->-0·5
-0·5->-1·0
-1·0->-1·5
-1·5->-2·0
-2·0->-2·5
-2·5->-3·0
-3·0->-3·5
-3·5->-4·0
-4·0->-4·5
-4·5>-5·0
29 30 31 1 2 3 4 5 6 7 8 9
JULY
AUGUST
SNOW TEMPERATURE (°C)
0·0->-0·5
-0·5->-1·0
-1·0-> -1·5
29 30 31 1 2 3 4 5 6 7 8 9
JULY
AUGUST

greatest growth must have occurred in the four days August 6-9. There is also the possibility that the temperature gradient through the depth hoar layer may be underestimated by the thermistor gradient. The depth hoar layer was only 150 mm thick while the thermistors covered a range of 300 mm and did not directly coincide with the location of the TG stratum. A discrepancy between the two gradients is supported by the temperature trace in figure 4.10. The temperature gradient in the depth hoar layer on August 10 was found to be twice that for the gradient between 50 and 300 mm above the ground surface. This relationship suggests that the temperature gradient in the depth hoar layer may have been as high as $30^{o}C\ m^{-1}$ during the period August 6 - 9. The validity of this assumption cannot be tested using these results alone. However, in view of the ten days required for the development of even early TG crystals with a gradient of $5\ mb\ m^{-1}$, it seems reasonable that much stronger gradients must have existed in the basal layers of the snowpack.

This example demonstrates how shallow snowcover and cold temperatures in the early or mid-winter combine to promote the development of depth hoar crystals. Although the indirect meteorological evidence indicated critical mean snowpack temperature gradients were not sustained long enough to produce early TG crystals, this analysis suggests that variations of gradients within the snowpack are adequate for even advanced depth hoar. The full role of variations in temperature gradients is discussed in the following section.

4.4.3.3 Variation in the temperature gradients

Variations in the temperature gradient within a snowpack occur because of two main reasons: differences in thermal conductivity of snow and the uneven nature of energy exchanges. The greatest energy exchanges occur at the surface of a snowpack. Hence, the surface snow layers experience the greatest extremes and fluctuations in temperature. The strongest temperature gradients occur near the surface because of cooling of the surface especially through radiative exchanges on cold clear nights. Over the long term, such gradients are eliminated by heat inputs to the surface (La Chapelle and Armstrong 1977). Fluctuations in the surface gradient occurs most regularly on a daily basis following the net radiation cycle. However, it was observed in the study area that when cold clear conditions persisted for a number of days, a strong temperature gradient prevailed at the surface of the snowpack, especially on shady south facing slopes. Frequently in these layers, early stages of TG metamorphism were apparent and in some cases fully developed hoar crystals were produced. Figure 4.12 illustrates the development of such a layer on a south facing slope at the end of June, 1980. The vapour pressure gradient through the stratum at the time of the snow pit was 9.8 mb m^{-1} ($27^{o}C\ m^{-1}$), and based on equation 4.4, only 5 days would be required for early TG crystal formation. Cold clear periods which would sustain this gradient are known to frequently occur for 5 days or longer on the Craigieburn Range.

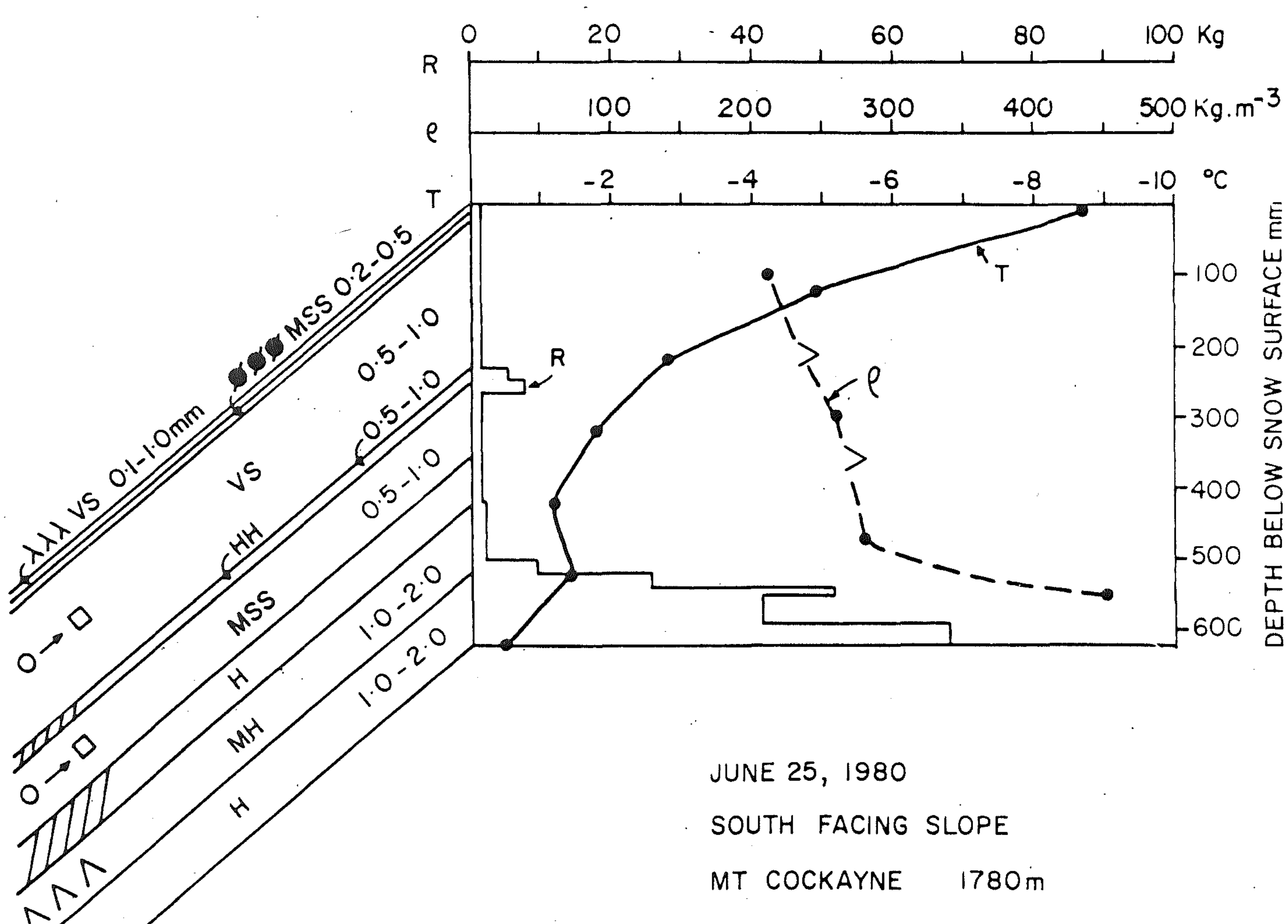


Figure 4.12 SNOW PIT PROFILE JUNE 25 1980.
Note the development of TG crystals above and below both ice crusts and immediately below the surface.

Strong variations in snow temperature gradients were also observed in other upper levels of the snowpack and frequently in association with ice crusts. Figure 4.12, in addition to containing the previously described surface TG crystals, also depicts reasonably well developed hoar crystals below an ice crust and early TG crystals above the same crust. This type of location is not uncommon for TG metamorphism, in fact a majority of all layers described in figure 4.9 were found in association with ice crusts or melt-freeze layers (table 4.4).

Despite the apparent importance of a TG-crust relationship, no published works exist dealing with scientific investigations of this phenomenon. The lack of study may be partially explained by the fact that the presence of crusts in the winter season is a much rarer phenomenon in colder continental snowpacks where a majority of the research on snowpack structure has been undertaken.

TYPE OF LAYER ABOVE TG STRATUM	%
SNOW COVER SURFACE	0
ET LAYER	22
ET-TG LAYER	13
CRUST OR MF LAYER	65
TYPE OF LAYER BELOW TG STRATUM	
GROUND SURFACE	29
ET LAYER	5
ET-TG LAYER	13
CRUST OR MF LAYER	53

Table 4.4 TYPES OF LAYERS ABOVE AND BELOW TG STRATA
The percentages refer to the frequency of observation for the fifty-five layers identified in figure 4.9.

It is hypothesized that TG metamorphism concentrates near crusts because disruptions in the temperature gradient and vapour transfer are created by the crust layers. The thermal conductivity of ice is known to vary appreciably from that of snow. For example Geiger (1961) notes that the thermal conductivity of snow with a density of 100 kg m^{-3} is approximately 0.03 W m^{-1} K^{-1} while that for snow-ice layers at 700 kg m^{-3} is 1.23 W m^{-1} K^{-1}. Hence, in the presence of ice and snow strata, appreciable thermal discontinuities would develop in the temperature gradient, specifically at the interface between different layers. The accuracy of dial stem thermometers usually employed in snow pit analysis precludes an accurate measurement of any such small scale irregularities which may occur between these strata. In order to overcome this problem, a DIGITRON digital probe thermometer (as described in chapter two) was used to measure snow temperatures in and around crusts. Although this research is only preliminary, some general observations may be made.

Figure 4.13 illustrates an example of an irregularity in the temperature gradient created by a thick ice crust. Over a large section of the snowpack, between 150 and 300 mm, the mean temperature gradient was $10^{o}C\ m^{-1}$ while the gradient about the ice crust was five times this value at $50^{o}C\ m^{-1}$. Within the temperature range illustrated in figure 4.13, this would equal a vapour pressure gradient of 20.6 mb m^{-1}, far in excess of the critical 5 mb m^{-1} separating ET and TG metamorphism.

Although temperatures could not be measured at any

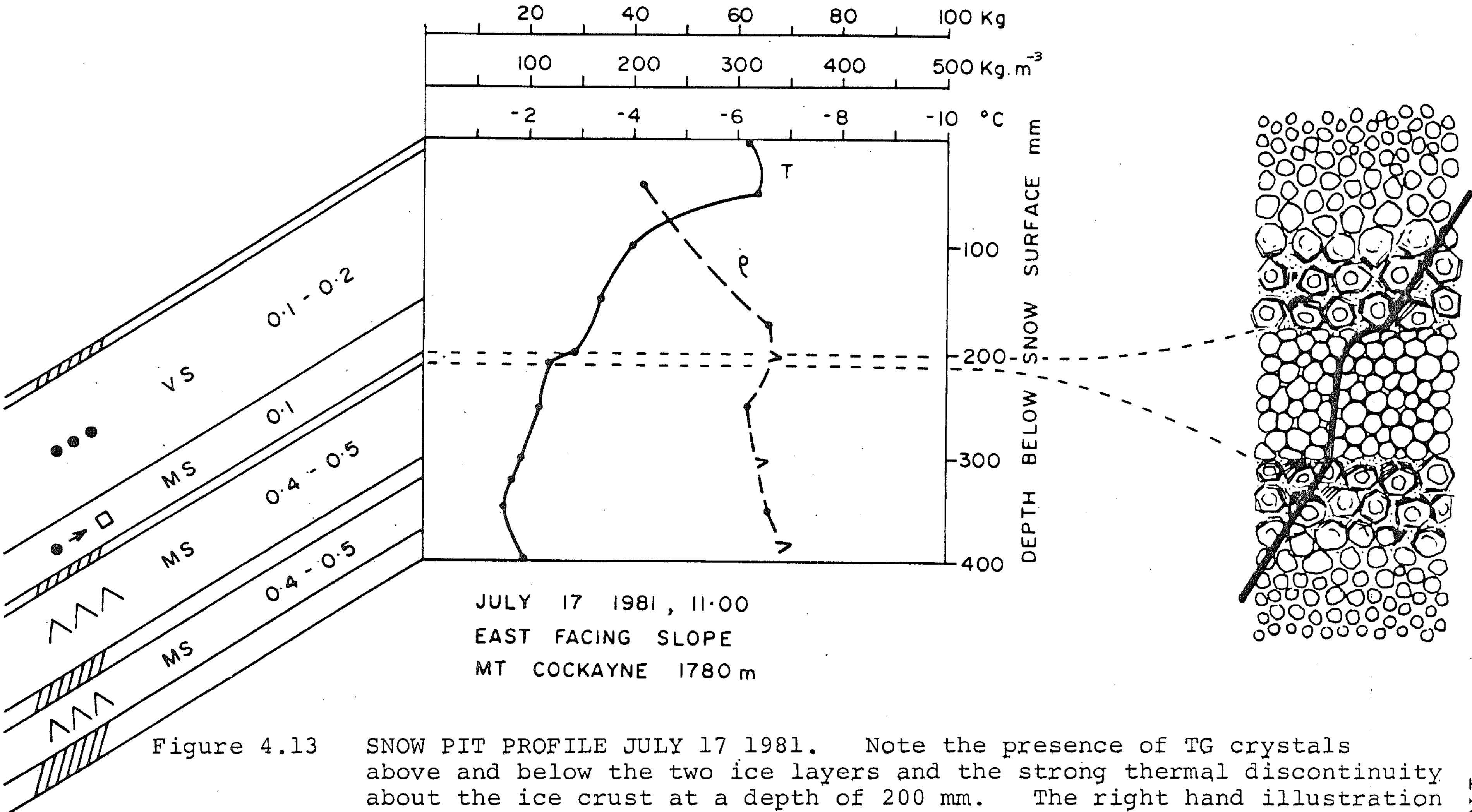


Figure 4.13 SNOW PIT PROFILE JULY 17 1981. Note the presence of TG crystals above and below the two ice layers and the strong thermal discontinuity about the ice crust at a depth of 200 mm. The right hand illustration depicts the temperature gradient through the thin ice layer and the growth of TG crystals above and below this stratum.

finer scale within the crust, the assumption was made that the greatest rates of change in temperature occurred near the surface of the crust. In view of the upward flow of heat and the high conductivity of ice compared to snow, this assumption seems reasonable. Therefore, with the crust being much warmer than the overlying snow, an appreciable vapour pressure gradient would exist across the ice crust-snow interface. The total supply of water vapour would also be great because of the concentrated mass of the ice layer. Hence, transfer of water vapour upwards from the ice crust into the overlying snow would produce TG crystals as illustrated in figure 4.13.

The presence of TG crystals below ice layers can be explained by a somewhat different process. The porosity and void ratio of snow is known to decrease with increasing density such that at 800 kg m^{-3} the matrix is termed 'ice' because it is no longer permeable (Mellor 1964). Although many crusts identified in a snowpack may not have a density as high as 800 kg m^{-3}, they are characterized by porosity values much lower than the surrounding lower density snow strata. Hence, they create an effective barrier to the transfer of water vapour. In the presence of a vapour pressure gradient, it is hypothesized that vapour will tend to accumulate beneath ice layers and sublimate onto existing grains forming TG crystals. It may also be that because of the restriction to vapour flow created by these ice crusts, a critical pressure gradient of 5 mb m^{-1} may not be necessary for the formation of TG crystals.

Much more research will have to be conducted to test the above theories but in view of the amount of field evidence as listed in table 4.4, ice crusts do play a significant role in TG crystal formation.

4.5 MELT-FREEZE METAMORPHISM

4.5.1 Background

Melt-freeze metamorphism occurs in a snowpack any time free water is produced and is subsequently refrozen within the ice matrix. Surface crusts are a product of melt-freeze metamorphism, the source of the free-water being either rain or meltwater and the refreezing due to nocturnal radiative cooling, colder air temperatures or because of cold underlying snow. Melt-freeze metamorphism may occur throughout entire snowpacks if sufficient free water is present. In cold snowpacks, MF metamorphism during the winter months is primarily limited to the development of surface crusts. Melting and refreezing of large portions of the snow column are confined more to the spring and summer months. Within warmer and more maritime climates, extensive MF metamorphism may occur at any time during the winter season.

Surface crusts resulting from melting and refreezing have been traditionally classified into two broad categories: rain and sun crusts. The first is obviously a product of rainfall and in a cold snowpack under conditions of heavy rain may develop to a considerable thickness, frequently in the order of 10 mm. Sun crusts are usually much thinner, especially during the winter months when radiation receipts are quite low. The term sun crust is also somewhat of a misnomer for regions in which radiation is not the dominant heatflow during non-precipitation melt periods. As will be pointed out in chapter six, this is the case, especially during the winter, for the Craigieburn Range.

The snow structure which is produced by melt-freeze metamorphism in many ways resembles that of equi-temperature metamorphism. In general, small ice grains are eliminated at the expense of larger ones, which eventually grow to 1.0 - 2.0 mm in diameter [Colbeck (1975a); Raymond and Tusima (1979); Wakahama (1968, 1975)]. Melt-freeze metamorphism may occur under two modes of free water saturation which Colbeck (1975b) has identified as funicular and pendular. Within the pendular mode, water saturation of the snowpack is low, occupying less than 14% of the total pore volume. Such low levels of saturation allow only a slow growth of large crystals and because inter-granular bonding remains intact, no increase in overall density occurs other than that due to the additional water. The strength of the snowpack may in fact increase because of the increased water tension between grains.

The funicular mode is associated with water saturation levels exceeding 14% of the pore volume. Growth rates are high (Wakahama 1968) and melting may occur at intergranular contacts resulting in a decrease of mechanical strength. As melting occurs at contact points, grains may also resettle producing an overall increase in density. Homogeneous snow under normal gravity drainage is characterized by water saturation values in the pendular range, but over impermeable layers, such as buried ice crusts, funicular states of saturation often develop.

4.5.2 Observations in the Craigieburn Range

Surface crusts were observed to be a regular feature of even the mid-winter Craigieburn snowpack. As might be expected from the frequency of winter rain events outlined in chapter three, many were rain crusts. Strictly speaking, most crusts which formed during non-rain periods could not be classed as sun crusts as they formed primarily from heat supplied by warm winds. Any crusts which did develop during the main winter season from primarily radiation heatflow were thin, commonly less than approximately 2 - 3 mm. Although such layers were difficult to trace once buried by subsequent snow, it is believed that because of the warm snowpack temperatures and the thinness of the crusts, most would rapidly disintegrate (Langham 1974).

The ice crusts which seemed to dominate the snowpack stratigraphy were those formed during the rapid and extensive melt periods which are described in chapter six. In fact during many of these events, large portions of the upper snowpack were inundated with either rain or meltwater and underwent melt-freeze metamorphism. Figures 4.14 and 4.15 illustrate the before and after snowpack stratigraphy of one such event. As illustrated in figure 4.14, the three layers between 700 and 1000 mm have been inundated with rain and meltwater, and covered by subsequent snowfalls.

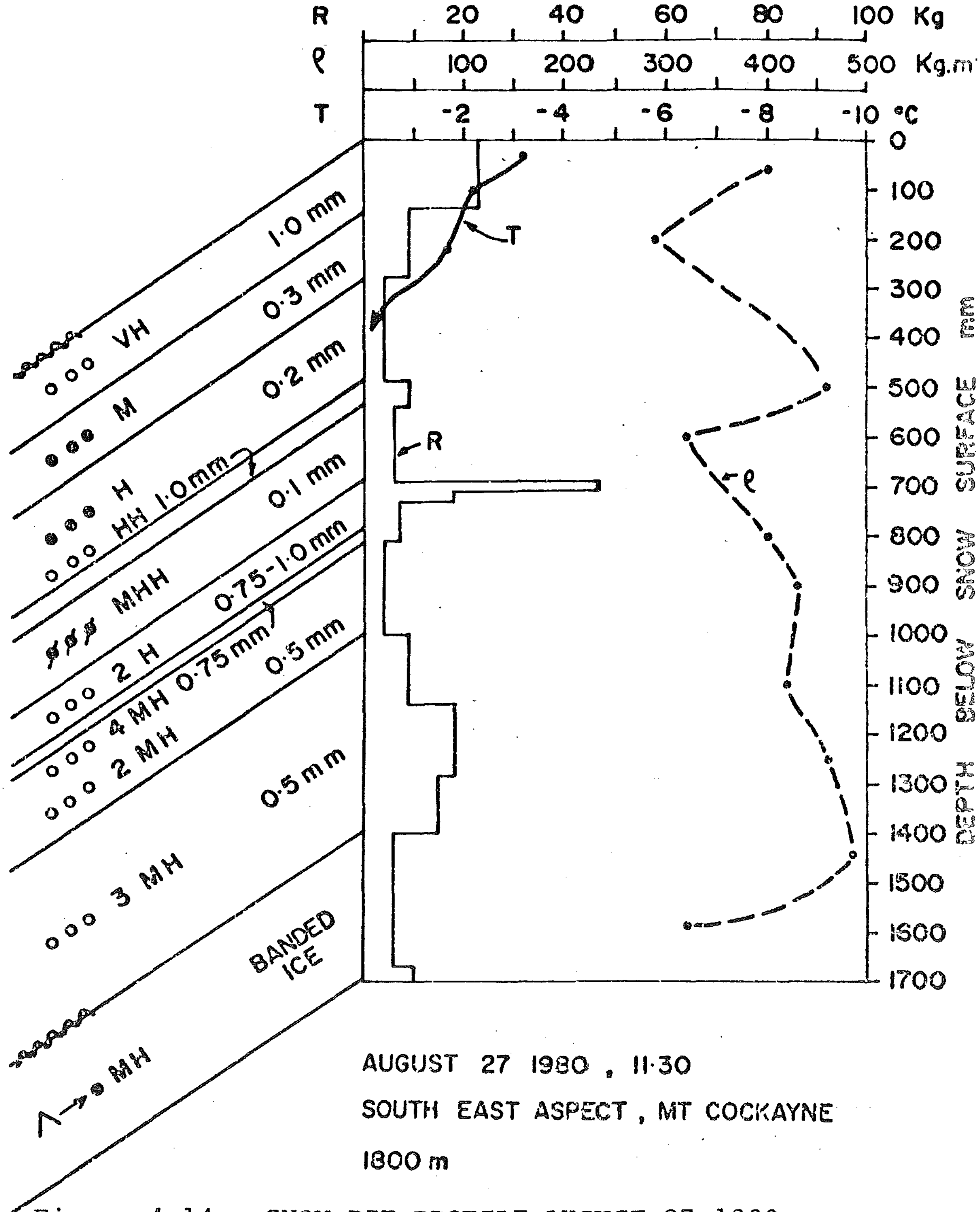


Figure 4.14 SNOW PIT PROFILE AUGUST 27 1980. A snowpack after appreciable warming from a North-westerly storm. Note the concentration of water (2, 4, 2) at approximately 800 mm deep within the snow column.

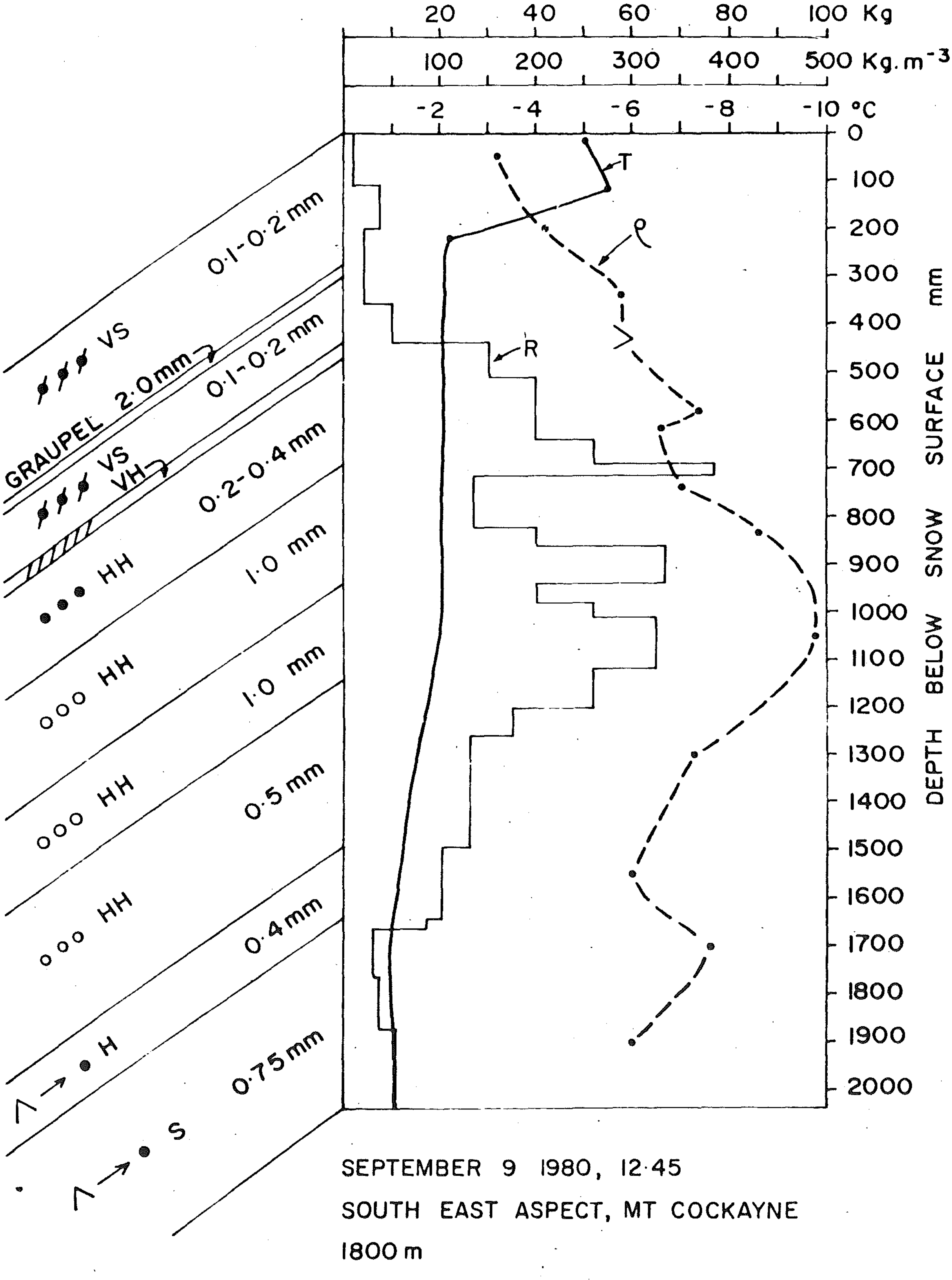


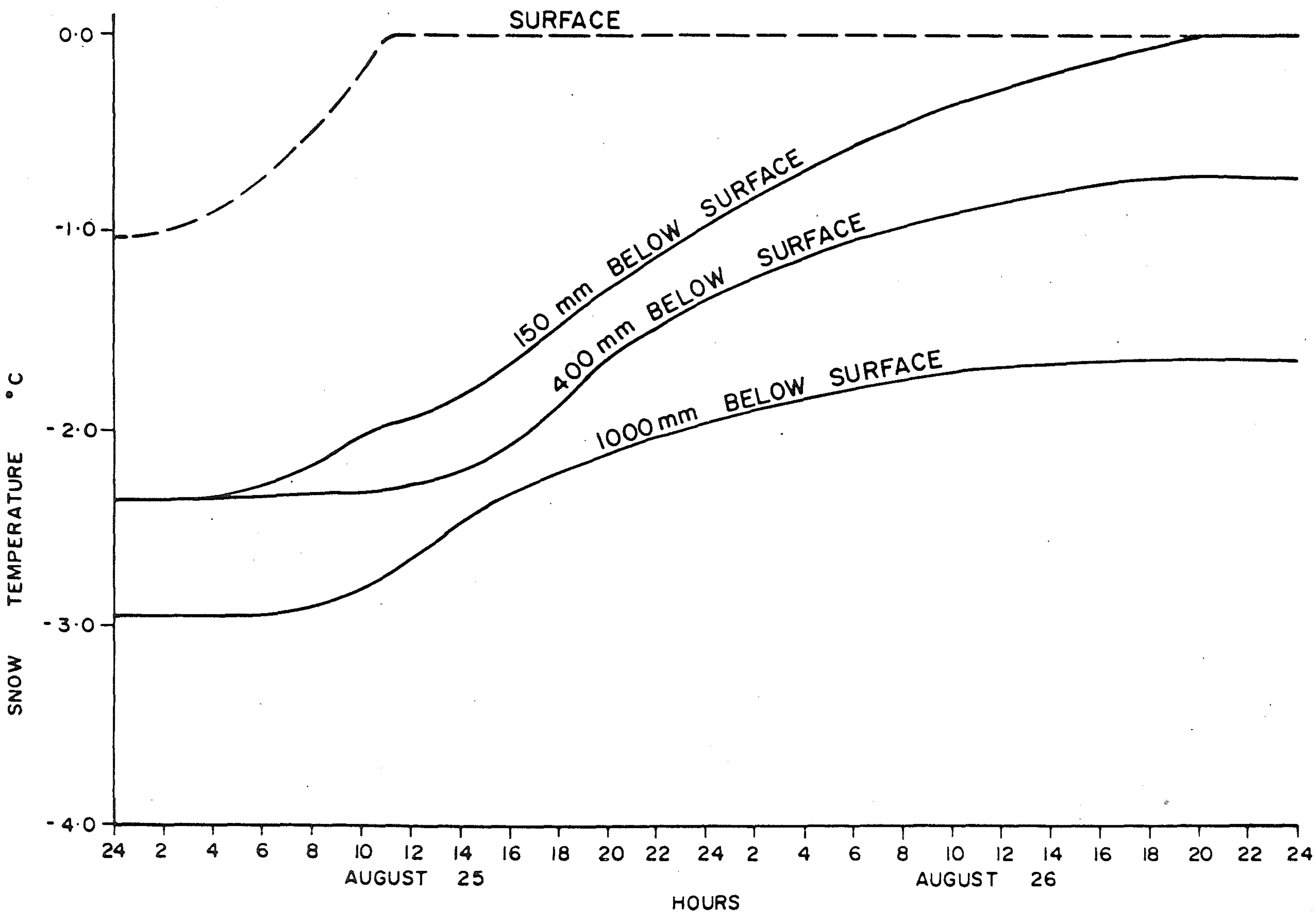
Figure 4.15 SNOW PIT PROFILE SEPTEMBER 9 1980. Note the considerable increase in mechanical strength since August 27 and the relative uniform nature of the middle layers created by MF metamorphism.

Although the layers have experienced appreciable ET metamorphism, as evinced by the large grain size and previous snow pit observations, the mechanical strength of the strata is quite low. Only the surface of the upper saturated layer has a ram value in excess of 20 kg. Even this value is likely the result of refreezing as the snowpack begins to cool down from an isothermal state at $0^{o}C$ produced by the rain and meltwater.

The greatest concentration of free water was found ponded in a layer 20 mm thick located between the two other saturated layers. Gravity drainage was obviously impaired at the base of this layer and saturation levels could be considered to be well within the pendular mode. The density of the layer, comprising both water and ice, could not be accurately measured because much of the free water drained from the sampler prior to weighing. A shovel test showed that this stratum was a definite sliding layer which is consistent with the theory that intergranular bonds break down under high levels of saturation (Colbeck 1975a).

Figure 4.15 illustrates the snowpack two weeks later on September 9, 1980, after a considerable number of changes had occurred, including the re-establishment of cold temperatures throughout the entire column. The layers which had previously been saturated are now totally frozen and exhibit relatively high ram values largely because of the free water bonding grains together. This type of bonding is different from that formed under ET metamorphism because entire grains become cemented together rather than simply being joined at inter-granular contacts. Grain size has also increased in these layers to approximately 1.0 mm along with a general rise in the density values.

The above example demonstrates the effect of MF metamorphism on snowpack structure when water is spread throughout the snow. However, in many cases the presence of crusts in the Craigieburn snowpack totally restricts the flow of water. Gerdel (19 54) noted that ponding of water is only a transient stage because ice planes in wet snow rapidly disintegrate. This case may be true for a melting spring snowpack but during the winter in the Craigieburn snowpack, refreezing of ponded water frequently occurred before the disintegration of underlying crusts could take place. Rapid changes in air temperature with frontal systems undoubtedly are accompanied by significant changes in snowpack temperature. Warm temperatures and rain are normally associated with north-westerly airflows (as established in chapters three and six) and cause rapid warming of the snowpack. Figure 4.16 illustrates the rapid warming associated with the passage of the north-westerly storm which occurred one day prior to the snow pit observations in figure 4.14. Although there are differences in the degree of snowpack warming (possibly due to instrument accuracy of the distance thermograph), between figures 4.14 and 4.16, the temperature traces outline the effectiveness of such storms in warming the snow, even at a depth of 1 m. When such events are quickly followed by a cold southerly airstream, the snowpack will rapidly cool and refreeze any ponded water. Frequent observations showed that free water repeatedly accumulated and refroze


SURFACE
150 mm BELOW SURFACE
400 mm BELOW SURFACE
1000 mm BELOW SURFACE
SNOW TEMPERATURE °C
0·0
-1·0
-2·0
-3·0
-4·0
24 2 4 6 8 10 12 14 16 18 20 22 24 2 4 6 8 10 12 14 16 18 20 22 24
AUGUST 25
AUGUST 26
HOURS

over particular ice layers. Such a layer was observed in the snowpack during August 1979 (figure 4.17). The layer, located approximately one metre from the surface, was 80 mm thick and in parts was totally comprised of blue or clear ice. An accurate density measurement could not be made with standard snow sampling equipment, but white ice on lakes, produced by the slushing of snowcover, is known to have an identical form to that of the observed ice layer (Adams and Prowse 1981) and has a density range of approximately 750 - 850 kg m^{-3}. The mechanical strength of the layer was also found to be exceedingly high, ranging in ram numbers from 152 to 2,000 kg, although the extreme values are thought to be inaccurate because of limitations in the ram for measuring highly resistant surfaces. However, they do demonstrate the abnormally hard nature of such layers for a snowpack.

An additional point of interest in figure 4.17 is the presence of large depth hoar crystals, characterized by very low mechanical strength and located directly below the ice layer. The slope on which this snow pit was observed, avalanched four times during the 1979 winter season (figure 4.18). The periodic removal of upper snow layers by avalanching may have assisted in the development of strong temperature gradients and the subsequent growth of depth hoar in the remaining snow. Similarly, the growth of the ice layer would have been enhanced if it had been directly exposed to rainfall and surface melt. Both the depth hoar and the surface of the ice layer are known to be good sliding layers for avalanches. The implications of this are beyond

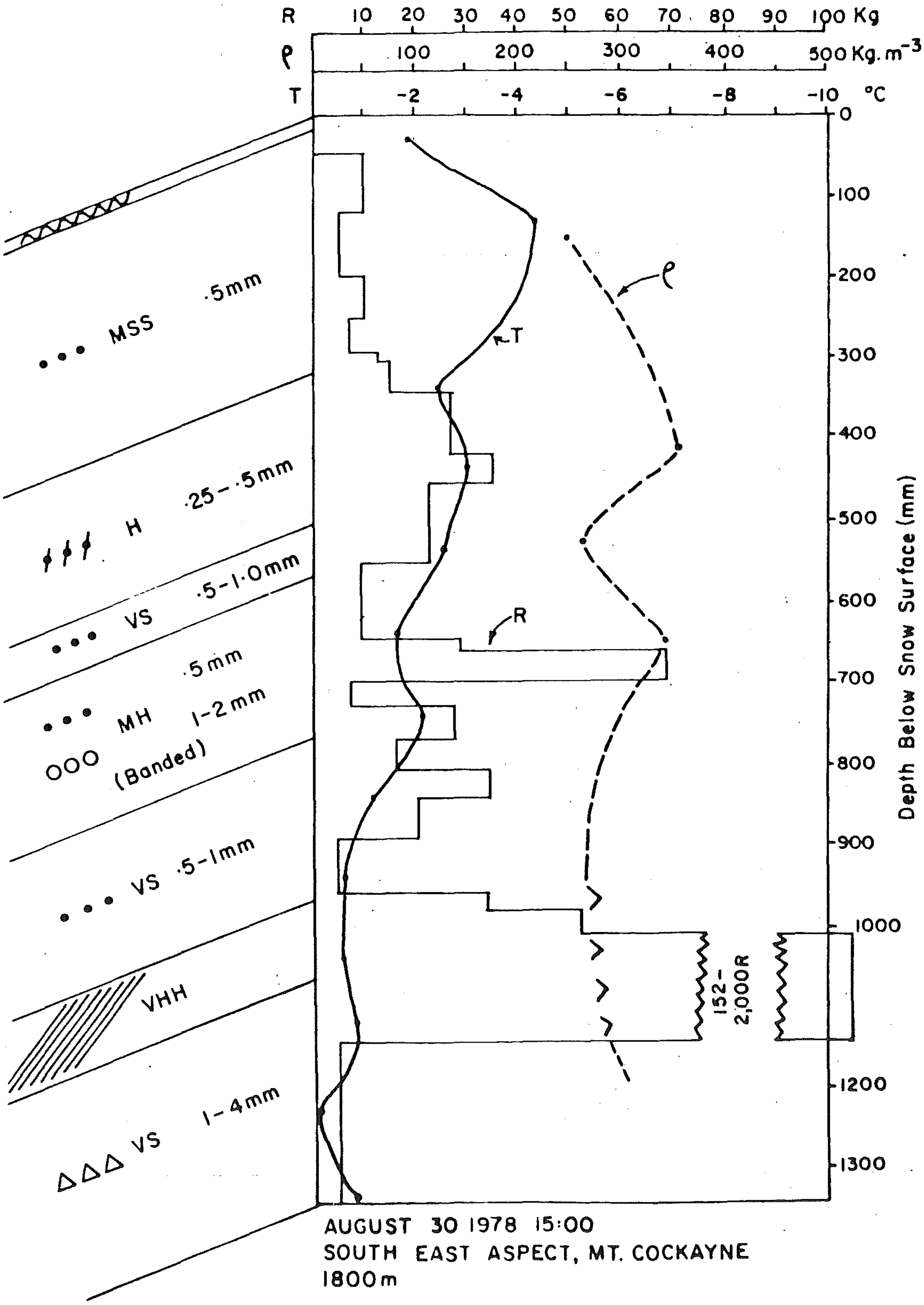


Figure 4.17 SNOW PIT PROFILE AUGUST 30 1978. Note the high ram numbers for the dense ice layer near the base of the snowpack and the underlying well developed depth hoar 1 - 4 mm in diameter.

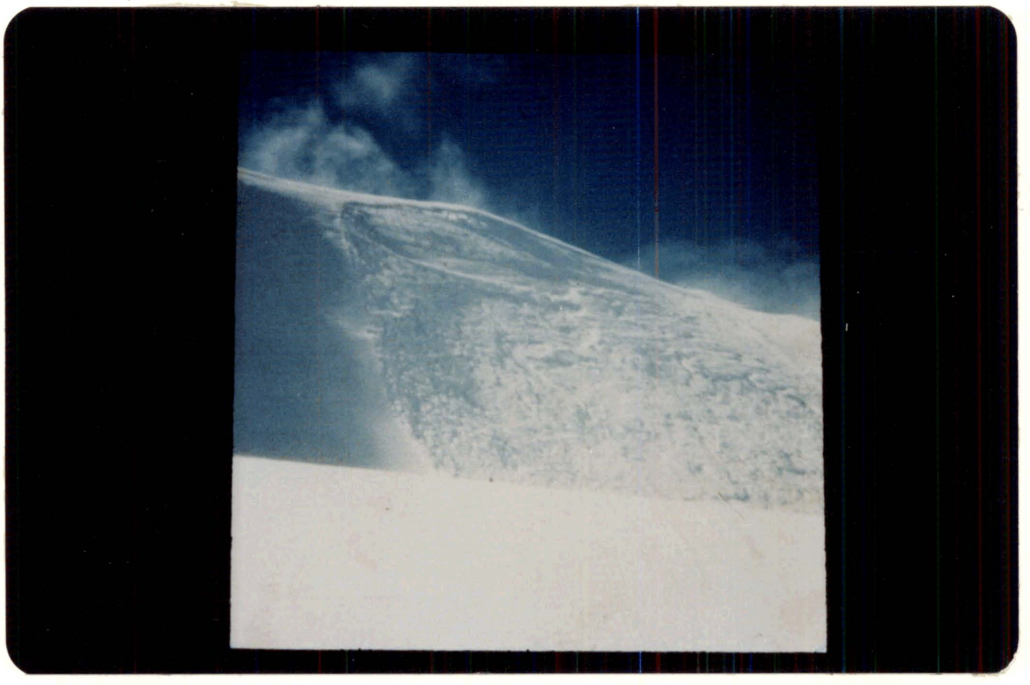


Figure 4.18 SOFT SLAB AVALANCHE MT COCKAYNE JULY 29 1979. Fracture line is approximately 120 m wide and 0.2 m deep and accumulated to 2 m in depth. Sliding layer was icy smooth surface. This was the fourth avalanche in the same location during the 1979 winter. Note the strong wind redistribution of snow at the ridge.

the scope of this report but are being considered by other researchers.

4.6 CHANGES IN MEAN SNOWPACK DENSITY

As a result of wind compaction and the three metamorphic processes outlined, the overall mean density of a snowpack should increase over time, interrupted only by the inputs of new light density snow. The rate of densification should also be a direct function of the mean temperature of a climate because rates of vapour transfer for both ET and TG metamorphism and the frequency of MF metamorphism are inversely proportional to the coldness of a snowpack. Hence, the slowest density changes are in polar, high alpine and cold continental snowpacks. The highest rates are in warm snowpacks, such as found in temperate maritime locations.

Figure 4.19 illustrates the changes in the mean snow density over the 1980 winter season at the SB site. Each daily value is the product of a 5 day binomially filtered mean based on approximately 580 density measurements taken at 100 - 200 mm depth intervals through the snowpack. Consistent with the relatively warm climate of the Craigieburn Range, rapid fluctuations ranging from approximately 10 - 25 kg m^{-3} d^{-1}, are prevalent throughout the main winter period, June to September. The full range of the fluctuations is in excess of 200 kg m^{-3}, from as low as 250 kg m^{-3} to over 450 kg m^{-3}. Most of the periods with densities in excess of 400 kg m^{-3} were associated with snowmelt conditions and MF metamorphism.

The greatest variations in snow density appear to occur in June and July as evinced by the coefficients of variation in table 4.5. As the season progresses, the mean monthly snow density increases by approximately 40 kg m^{-3} mh^{-1}, but variations in the daily mean values decrease such that by October the coefficient of variation is only 10%.

The greater consistency of density values in the later months reflects the reduction in inputs of new light

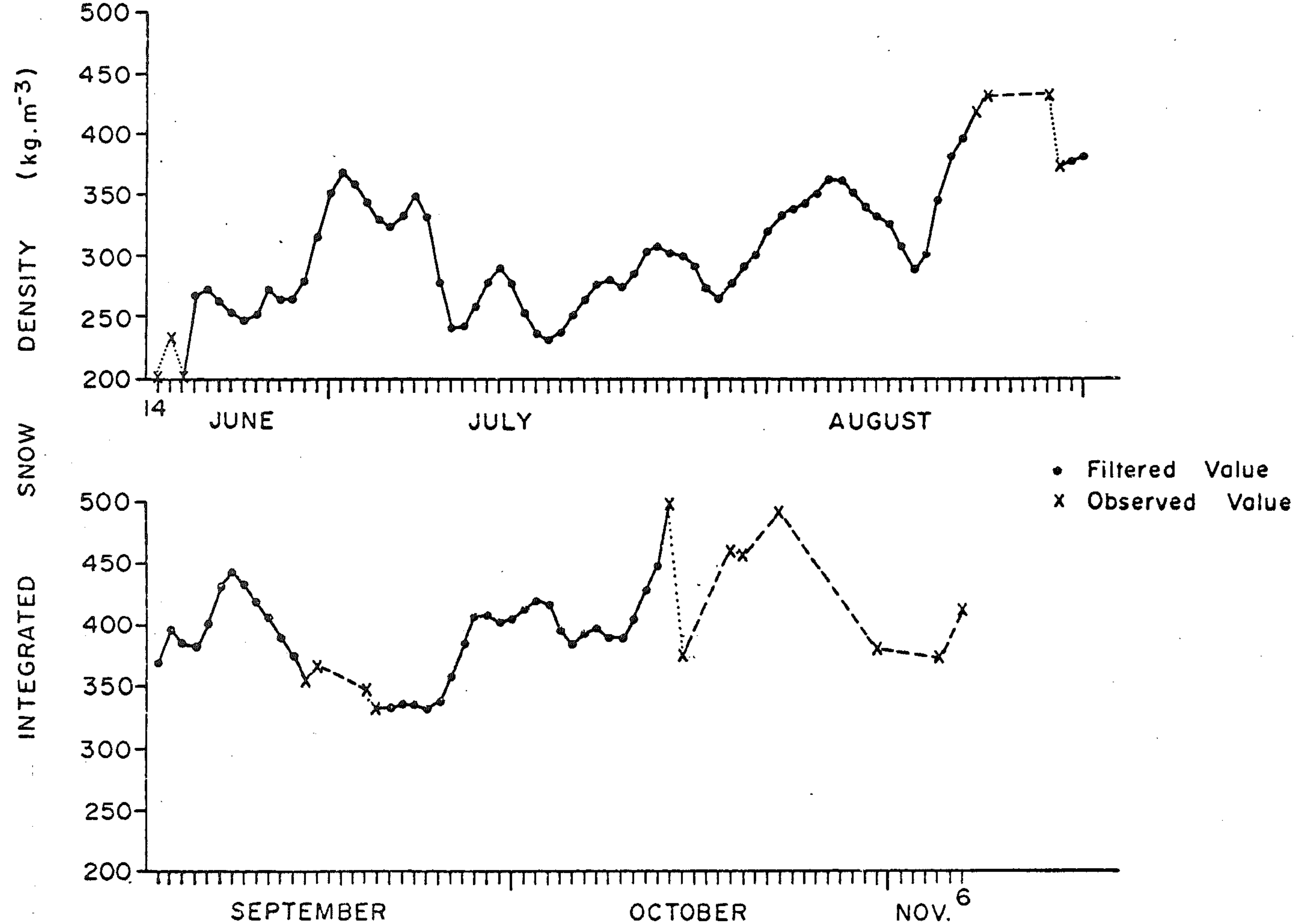


Figure 4.19 TEMPORAL TRENDS IN MEAN INTEGRATED SNOW DENSITY. Data are based on density profiles conducted at the SB climate station. All filtered values were derived from a 5 day binomial filter.

	SAMPLE SIZE (DAYS)	MEAN DENSITY (kg m-3)	COEFFICIENT OF VARIATION (PERCENT)
JUNE	17	265	17
JULY	31	290	18
AUGUST	26	344	15
SEPTEMBER	21	383	13
OCTOBER	17	419	10
JUNE-SEPTEMBER AVERAGE		321	

Table 4.5 MEAN MONTHLY INTEGRATED SNOW DENSITY. Data is from daily snow density profiles conducted at the SB climate station over the 1980 winter season.

density snow and the relatively uniform density of melting snow. The mean monthly density in October was 419 kg m^{-3} but most late season melt phases were characterized by densities of approximately 450 kg m^{-3}. Such values agree with the generally held view (Smith 1974) that spring snow densities are commonly in the range 400 - 500 kg m^{-3}.

A number of researchers [for example: Bilello (1974); Dmitrieva (1954); U.S. Army (1956)] have attempted to relate the changes in mean snowpack density to climatic variables. The most common variables employed have been temperature and windspeed, variables which are also frequently correlated with density changes of specific layers from ET metamorphism.

Using similar methods a multiple regression analysis was attempted for the 1981 SB data based on daily changes in mean snow density, windspeed and air temperatures. Only periods dominated by rapid changes in mean snow density were analyzed and these were separated into early, mid and late season phases. The results of the regression appear in table 4.6. Temperature and windspeed proved to provide poor explanations for changes in snow density. Only the correlation coefficient for the early season, June-July period, exceeded 0.5. These poor correlations may be explained by the small sample sizes and difficulties in measuring temporal variations in mean snow density within alpine environments. In addition, the use of air temperatures and windspeed as the only meteorological indices may underestimate the role of the metamorphic changes, especially those due to melt-freeze metamorphism. Much more data will have to be collected concerning changes in snow density

PERIOD OF ANALYSIS	CORRELATION COEFFICIENT	REGRESSION EQUATION
11 DAYS JUNE-JULY	0.72	$\Delta\bar{\rho}_s = -17.65 + 1.57T_x + 0.83W$
16 DAYS AUGUST	0.15	$\Delta\bar{\rho}_s = 17.67 - 0.54T_x - 0.01W$
11 DAYS SEPTEMBER-OCTOBER	0.45	$\Delta\bar{\rho}_s = 37.69 - 1.21T_x - 0.04W$

Table 4.6 REGRESSION EQUATIONS FOR THE DAILY CHANGE IN MEAN SNOW DENSITY. $\Delta\bar{\rho}_s$ (kg m^{-3}) as explained by daily maximum air temperatures, T_x (^{o}C) and daily total windspeed W (km d^{-1}).

before a reliable relationship can be established. This may be possible in the near future as daily measurements of snowpack density have now been included as part of the standard weather observations at the SB climate station.

Although the above regression analysis proved unsatisfactory, a useful comparison can be made between the 1980 mean snow density values and the results of other researchers. Bilello (1974) separated North America into four density zones (table 4.7) which he also correlated with windspeed and air temperature. The density values used in the comparison were calculated over the November to March period which corresponds to the last entry of 321 kg m^{-3} in table 4.5 for the Craigieburn snowpack. This value is comparable to the Arctic and Sub-Arctic category for North America. However, in his analysis Bilello (1974) eliminated all months in which the mean air temperature

REGION	STRENGTH OF WIND	SNOW DENSITY (kg m^{-3})
INLAND	LIGHT	< 240
VARIABLE	MODERATE	≥ 240 - < 270
INLAND AND COASTAL	MODERATE TO STRONG	≥ 270 - < 310
ARCTIC AND SUBARCTIC	STRONG AND FRIGID	≥ 310

Table 4.7 SNOW REGIONS OF NORTH AMERICA. Regions are defined by Bilello (1974) on the basis of snow density and the strength of wind.

was above the freezing point. As noted in table 4.3 this proved to be the case for September 1980. Hence, based on only the months June to August the mean winter snow density for 1980 at the SB site would average 300 kg m^{-3} corresponding to Bilello's (1974) inland-coastal classification. This type of classification is further reinforced in figure 4.20, which considers temporal trends in density values for two locations in the Craigieburn Range and a number of places in North America. A strong similarity exists between the magnitude and rate of change of density from this study and those from Allan's Basin recorded by O'Loughlin (1969a), but the most useful comparisons can be made to the North American results. The snow densities at the SB site are consistently greater than those of the inland coastal and interior continental climates of North America as represented by the British Columbia Trench, Eastern Rockies and Northern Manitoba (McKay 1970). A density difference of approximately 100 kg m^{-3} separates the SB and B.C. Trench results. However, on average, the SB values are 80 kg m^{-3} less than those for the maritime snowpack as represented by the Mt Seymour, B.C. data (Fitzharris 1975). This middle position would support an inland-coastal classification for the snow densities of the Craigieburn Range. Surprisingly the data for the Arctic Archipeligo are also very similar to those of the SB site although the densification process in the Arctic is known to be much more related to wind compaction and less to the time dependent metamorphic processes described in this chapter.

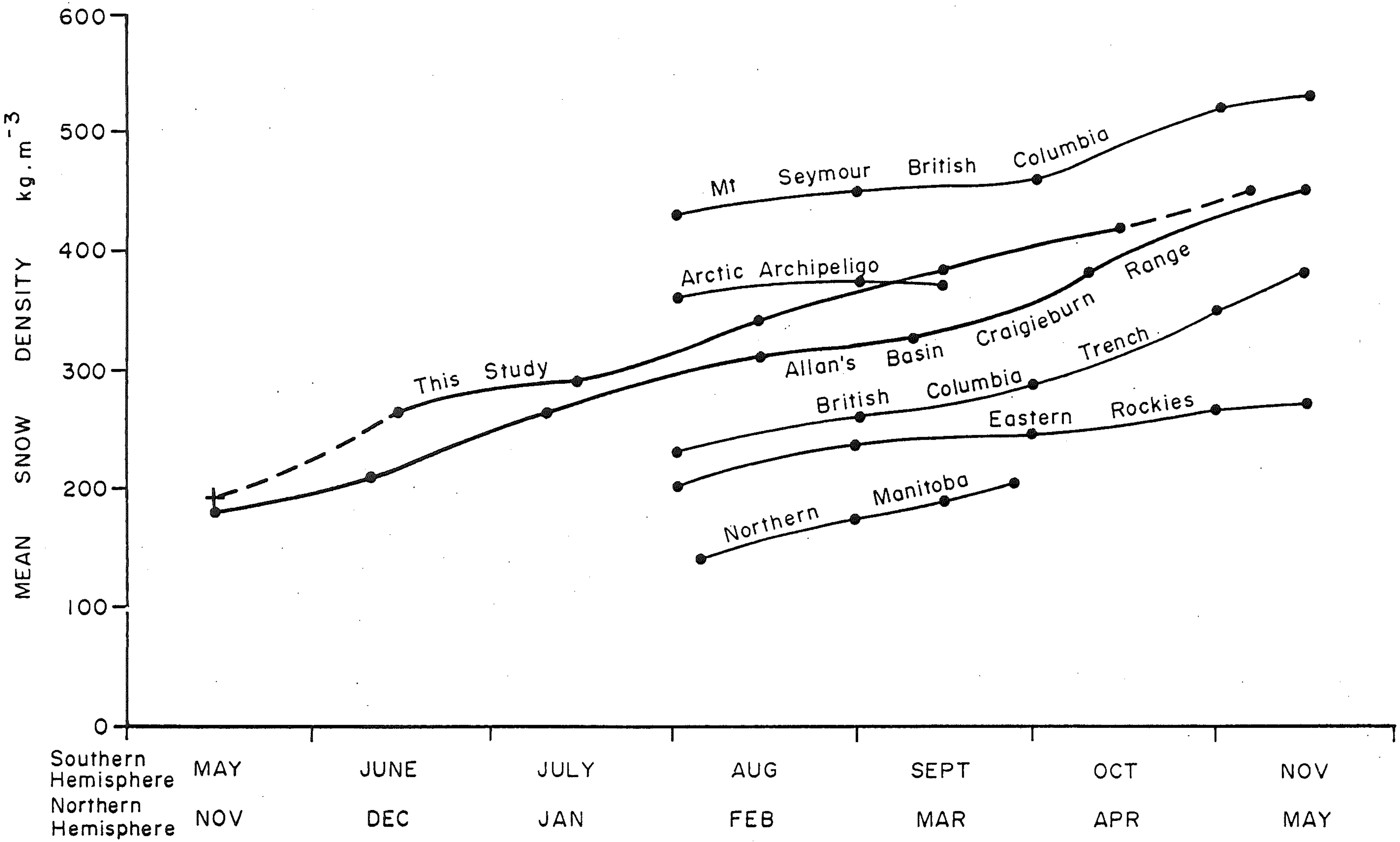


Figure 4.20 COMPARISON OF TRENDS IN MEAN SNOW DENSITY. The results from this study seem to be approximately mid-way between maritime and continental snowpacks of the Northern Hemisphere.

4.7 SUMMARY

The overall structure of the Craigieburn snowcover has been demonstrated to be a product of a diverse range of climatic effects. Wind composition, large temperature fluctuations and frequent winter melt periods all combine to produce a highly complex, heterogeneous snowpack.

Wind, especially in the higher elevations, fragments and compacts much of the new snowfalls such that the upper layers are characterized by grain sizes and densities similar to that of snow at an intermediate stage of ET metamorphism. The mean density of wind compacted snow is 250 kg m^{-3}, over 120 kg m^{-3} greater than new snow which falls under calm conditions. Characteristics of the wind deposited snow conform with some of the definitions for soft and hard slab offered by other researchers, although no significant differences in density-strength relationships were found between these layers and ones formed primarily by ET metamorphism within the snowpack.

The most rapid rates of ET metamorphism were observed to occur in the new light density snow, resulting in density increases of approximately 50% in one day. In many cases advanced ET metamorphism was accelerated by the presence of large amounts of free water such that the effects of ET metamorphism could not be separated from those of melt-freeze metamorphism.

Frequent melt periods produced surface crusts but the magnitude of many of the events was such that the snow column became inundated with free water. During such melt phases significant decreases in the strength of the snow matrix were noted, especially where water ponded at semi-impermeable crusts. Rapid cooling usually followed melt periods and caused refreezing of the ponded water before disintegration of the flow impeding crusts could occur. Hence, ice layers of very high density and strength tended to develop with each new melt-freeze period.

In zones of the snowpack where water had freely drained, the growth of grain sizes was noticeably accelerated although no appreciable increase in density occurred other than that due to the presence of additional water.

Indirect climatic evidence suggested that TG metamorphism was likely to be rare in the Craigieburn snowpack. However, three locations were identified in the snow column at which TG crystals frequently developed. These included (a) a near surface TG crystal which formed primarily on south facing slopes during periods of strong negative radiation balance, (b) basal depth hoar which developed at the beginning of the snow season under shallow snow cover and (c) TG crystals which formed adjacent to ice crusts. The average density of well developed depth hoar layers was approximately 280 kg m^{-3} but in one particular case was as low as 180 kg m^{-3}. On average TG metamorphism affected 12% of the total snowpack thickness. The prevalence and frequently mature state of TG affected snow is surprising in view of the generally held view that depth hoar is primarily a product of cold continental climates.

The net effect of new snow inputs, wind compaction and metamorphic processes was to produce mean snowpack densities characterized by frequent and rapid fluctuations.

In general, the mean density increased by approximately 40 kg m^{-3} mth^{-1}, from 265 kg m^{-3} in June to well over 400 kg m^{-3} in October. Variations in density also decreased over time until a relatively uniform melting snow density of 450 kg m^{-3} was reached. Both the magnitude and trend in mean snow density could be considered representative of an inland-coastal classification, between those for continental and maritime climates.

CHAPTER V

BACKGROUND TO SNOWMELT

5.1 INTRODUCTION

This chapter reviews the theory and background of snowmelt production under three sections. The first section discusses the thermodynamic relationships between ice and water using the context of meltwater generation within a snow column as the focus. Specific attention is paid to the locations within a snowpack where changes of state are most likely to first occur and how subsequent processes serve to fully ripen the snow.

The second section details the individual components of the energy balance and focuses on measurement-calculation problems within an alpine environment. The background theory for many of the equations used in chapter six are also introduced.

The final section considers the relative importance of individual terms in the energy balance equation through a review of overseas research.

5.2 MELTWATER PRODUCTION

The melting of a snowpack may occur once the energy supply to the snow is sufficient to raise the temperature of the ice matrix to $0^{o}C$. Any further additions of heat will produce melting. The amount of energy required to raise the temperature of a given volume of ice is small compared to the amount of heat required to melt the same volume, specific heat of ice, C_i, (2.09×10^3 J kg^{-1})

being much smaller than the latent heat of fusion of ice, L_f (3.33×10^5 J kg^{-1}).

Energy transfers to a snowpack occur primarily at the snow-air interface. When there is a net energy gain by the snow surface which exceeds the rate at which this energy may be transferred to the remainder of the snowpack, surface warming and eventually melting will occur. Since the primary method of heat transfer within snow, in the absence of flowing water, is by the relatively slow process of conduction, surface melt frequently occurs before the entire snow column becomes isothermal at the melting point.

Unless evaporated away, water produced by surface melt will drain by gravity into the lower layers of the snowpack. The actual depth of penetration is dependent on the heat deficit or coldness of the underlying snow. The total volume of meltwater passing into a sub-freezing snow layer will refreeze if the heat deficit of that layer is equal to or exceeds the amount of latent heat which would be liberated during the refreezing process. The latent heat released in refreezing is then used to raise the temperature of the snow layer. If the heat deficit is insufficient to cause total refreezing of the meltwater, only part of the water will refreeze and the remainder will pass into a lower layer.

During the cold winter months the total heat deficit of a snowpack is at a maximum and hence most small amounts of meltwater will refreeze at the surface. Similarly, light rainfalls will freeze in the upper layers and produce a melt-freeze crust. During lengthy periods of high energy transfer and major rainfall events, liquid water may penetrate further into the snowpack and appreciably reduce the total heat deficit.

While a heat deficit remains in a snowpack, no appreciable runoff may be generated. Some water may be melted from the base of the snowpack but, because the energy transfers at the snow-ground interface are so small, this amount is not significant. Runoff can only occur once the entire heat deficit has been eliminated and the field capacity of the ice matrix has been satisfied. [The field capacity is usually below 7% of the total snow volume (Colbeck 1973)]. Further additions of energy to the snowpack will result in the production of meltwater and subsequent runoff. However, as water is already held in the pore spaces by capillary pressure and surface tension, the amount of energy required to melt a given volume of snow is less than that for pure ice.

The ratio of the amount of heat required to produce a given volume of water from snow to melt the same volume from pure ice at $0^{o}C$ is referred to as the thermal quality (U.S. Army 1956). It has also been referred to as the ratio of the weight of ice to the total weight of a unit volume of snow [Bruce and Clark (1966); Dunne and Leopold (1978)]. For a melting snowcover the thermal quality has been found to vary from 0.90 to the maximum value of 1.00.

Snowmelt runoff may occur from snowpacks during the winter season especially in the warmer and more maritime climates. However, winter melt periods are usually brief and once a negative energy balance becomes re-established over the surface of the snow, a heat deficit will redevelop throughout the snowpack. Addition of more snow will further increase the heat deficit.

5.3 THE ENERGY BALANCE

For seasonal snowcover, there is eventually a period when energy supplies to the snow are sufficient to cause total ablation. Although this is usually associated with the spring period, it may vary considerably in duration and time of year depending on the climate. Many of the terms in the energy balance equation may be seen to have distinct climatic variations. The validity of this observation is discussed in the last section of the chapter following descriptions of the individual energy balance components which may be written as:

$$Q_M = Q^* + Q_H + Q_E + Q_P + Q_G - Q_\Theta \qquad (5.1)$$

where Q_M is the heat available for melt; Q^* is the radiation heat flow; Q_H is the sensible heat flow; Q_E is the latent heat flow; Q_P is the precipitation heat flow; Q_G is the heat flow from the layers underlying the snow; Q_Θ is the heat deficit which for a melting snowpack can be treated as 0.0. All are in units of W m^{-2}.

5.3.1 Radiation Heat Flow

The net radiation balance for a snowcover can be written as:

$$Q^* = K^* + L^* = K{\downarrow}\ (1 - a_k) + L{\downarrow} - L{\uparrow} \qquad (5.2)$$

where K^* is the net shortwave radiation balance 0.3 - 3.0 um; L^* is the net longwave radiation balance termed effective radiation 3.0 - 80 um; $K{\downarrow}$ is the combined direct and diffuse solar radiation, termed global radiation; $L{\downarrow}$ is the longwave atmospheric radiation; $L{\uparrow}$ is the long wave terrestrial radiation; (All units in Wm^{-2}) a_k is the surface snow albedo for shortwave radiation.

For many snowmelt investigations, it has not been practical to measure all the components of the radiation balance. Even when this has been possible, single measurements in mountainous terrain are of only limited use for large scale applications. The following discussion reviews some of the methods for calculating radiation and some of the problems to be encountered in alpine topography.

5.3.1.1 Shortwave Radiation

The amount of global radiation reaching the surface of the earth has been thoroughly examined [for example: Geiger (1961); Kondratyev (1969); Garnier and Ohmura (1968)]. In general, direct solar radiation reaching a horizontal surface can be expressed as a function of the solar constant, latitude, time of day and year, and prevailing atmospheric conditions. As most surfaces are not horizontal nor at a constant elevation, adjustments must be made for surface slope and the thickness of the optical air mass. In general, the greatest receipts of shortwave radiation are on north facing slopes in the southern hemisphere. Under clear skies, direct solar radiation should also increase with elevation because of decreases in the thickness of the optical air mass and increases in atmospheric transmissivity (Kuzmin 1961). Garnier and Ohmura (1968) offer one of the simplest methods for estimating incident direct beam radiation at the surface of the earth. A thorough discussion of the methods is contained in Obled and Harder (1979).

Within areas of extensive relief, the calculation of direct beam radiation is further complicated by shadowing from surrounding topographical features. Digital terrain models as developed by Dozier (1979) and Williams et al. (1972) can account for such irregularities.

The calculation of diffuse radiation has been found to be more complex than that for direct beam radiation. The amount reaching a horizontal surface can be approximated by considering the solar angle, concentration of aerosol particles and air density (Dozier 1979) and as for direct radiation, estimates for sloping surfaces can also be made (Kondratyev 1969). Within alpine areas, the portion of the sky dome emitting diffuse ratiation may also be severely restricted. However, additional diffuse radiation may be received from solar radiation reflected from adjacent slopes or from obscured areas where it is redirected back down by scattering and reflection in the atmosphere. Kuzmin (1961) has suggested that such contributions can actually increase the total amount of scattered radiation received at a snow surface compared to bare ground.

Another problem with estimating diffuse radiation, especially for clear sky conditions, is that it is anisotropic (Kondratyev 1969) and not isotropic as assumed in most calculations. Intensities are greatest in the vicinity of the sun and at the horizon.

Important for mountainous areas is the fact that the quantity of diffuse radiation decreases with elevation because of the decreasing optical air mass and atmospheric turbidity. Obled and Harder (1979) offer a thorough review of the problems associated with estimating diffuse radiation and Dozier (1979) presents a computer model for its estimation.

5.3.1.2 Albedo

The amount of radiation actually entering a snowpack is dependent, as outlined in equation 5.2, on the albedo of the snow surface. Snow is known to act like an almost perfect black-body, reflecting less than 0.5% of incident longwave radiation (Geiger 1961). In contrast, the albedo to shortwave radiation may be quite large. Maguire (1975) reports a range in a_k from as little as 0.25 (25%) for old melting snow to as high as 0.95 (95%) for fresh new snow.

Work by various investigators [Bergen (1975); Bohren and Barkstrom (1974); Gaitskhoki (1971); Kuznetsov (1960); Thomas (1962)] have demonstrated albedo for snow and ice to be a function of numerous parameters including surface irregularities, grain size, grain orientation, density, porosity, water content and impurities. In general, decreases in the snow albedo for melting snow occur because of a decrease in porosity and increases in grain size, density, free water content and impurities.

In the absence of direct measurement of radiation, albedo values must be approximated. The most common method is to use a relationship already established between albedo and a surface snow property or meteorological index, usually related to melt production. Anderson (1976) established an albedo relationship using surface snow density while the U.S. Army (1956) has employed accumulated degrees or 'time since last snowfall'.

Albedo values used in snow research are normally integrated for the full shortwave spectrum. However, the reflection powers of snow vary according to wavelength and as the spectral composition of solar radiation changes during the day, so do radiation receipts at the snow surface [Kalitin (1930); Kondratyev (1969)]. As pointed out by Greenbank (1945) and Gaitskhoki (1971) the spectral differences seem to be of such a small order of magnitude that snow can almost be considered unselective, except in the longer wavelengths (O'Brien and Munis 1975).

Albedo has also been found to vary in regards to direct and diffuse shortwave radiation. Snow reflects diffuse radiation equally at all angles of incidence but for direct beam radiation the intensity of reflection increases with the angle of incidence (Kondratyev 1969).

5.3.1.3 Longwave Radiation

For many snowmelt applications, effective longwave radiation is estimated based primarily on the Stefan-Boltzman constant. The amount of terrestial radiation emitted by a snow surface can be calculated according to:

$$L\uparrow = \varepsilon_s \sigma T_s^4 \qquad (5.3)$$

where ε_s is the emissivity of snow; σ is the Stefan-Boltzman constant (5.6697×10^{-8} W m^{-2} K^{-4}); T_s is the snow surface temperature (K).

The emissivity of snow is very high and for practical applications is normally set to a value between 0.95 and 1.0 [for example: 0.99 by Anderson (1976); 0.98 by Matveev (1965); 0.95 by Marks (1979)].

During sub-freezing conditions the amount of longwave radiation emitted by snow will decrease with elevation just as the snow surface temperature decreases. However, during snowmelt conditions the surface temperature cannot rise above 0^{o}C and the value of $L\uparrow$ remains constant.

The equation for atmospheric longwave radiation is similar to equation 5.3 but the emissivity value for the atmosphere is derived from near surface moisture and temperature measurements. In the general form, the equation for clear sky conditions may be written as:

$$L_o\downarrow = \varepsilon_a \sigma T_a^4 \qquad (5.4)$$

where $L_o\downarrow$ is the amount of longwave radiation emitted from clear skies ($W\,m^{-2}$); T_a is the air temperature (K); ε_a is the effective atmospheric emissivity.

A number of researchers have produced variations of equation 5.4. One of the most commonly used equations was derived by Brunt (1932) which requires measurements of near surface air temperature and moisture levels. Other equations have also been employed which assume constant moisture levels and hence air temperature is the only required data input [for example: U.S. Army (1956) developed after Brunt (1932), and Monteith (1973) after Swinbank (1951)]. The U.S. Army (1956) reasoned that for snowmelt conditions atmospheric moisture levels would be high and relatively constant. However, many snowmelt periods are characterized by quite low humidities, as experienced during föhn or Chinook conditions [Boyd (1967); Golding (1978)]. A comparison of the Monteith (1973) and U.S. Army (1956) equations was made to the more recently developed Brutsaert (1975) equation (appendix G). The results of the comparison demonstrate the insensitive nature of the purely temperature dependent equations and the value of the Brutsaert (1975) formula for predicting $L\downarrow$ under a range of humidity conditions.

The Brutsaert (1975) equation, as the original Brunt (1932) equation, requires data inputs concerning air temperature and vapour pressure but does not require site specific empirical parameters. Marks(1979) attests to its usefulness in alpine snowmelt computations.

According to Brutsaert (1975) the formula for the effective atmospheric emissivity in equation 5.4 is:

$$\varepsilon_a = 1.24 \ (e_a/T_a)^{.14} \tag{5.5}$$

where e_a is the near surface vapour pressure (mb); T_a is the near surface air temperature (K).

As pointed out by Marks (1979) this equation was derived for sea level conditions and must be adjusted for use in alpine areas by correcting for barometric pressure and temperature differences with elevation. Therefore, according to Marks (1979):

$$\varepsilon_a = 1.24 \ (e'_a \ /T'_a \ )^{.14} \ (P_a/1013) \tag{5.6}$$

where P_a is the near surface air pressure (mb); T'_a is the adjusted sea level temperature (K); e'_a is the adjusted sea level vapour pressure (mb).

The values for corrected temperature and air pressure may be derived according to:

$$T'_a = T_a + (.0065z) \tag{5.7}$$

$$e'_a = (e_a/e_\upsilon)\ e'_\upsilon \tag{5.8}$$

where z is the elevation of the study site (m): e_υ is the saturation vapour pressure (mb) at temperature T_a; e'_υ is the saturation vapour pressure at temperature T'_a .

Hence, equation 5.4 may be rewritten as:

$$L_o\downarrow = 1.24\ (e'_a\ /T'_a\ )^{.14}\ \sigma\ T_a{}^4 \tag{5.9}$$

As described for shortwave radiation, alpine topography can also affect the receipts of effective radiation. The amount of atmospheric radiation is controlled by the size of the sky dome but at low level locations any reduction may be compensated by terresial radiation emitted from the obstructing topographical features. The amount of compensated longwave radiation is usually greater than the lost atmospheric radiation, especially if some of the surrounding slopes are snow free. Hence, the amount of longwave radiation incident on a snow surface is likely to be greater for a valley rather than an open flat location.

The unobscured portion of the sky is referred to as the view factor and may be expressed according to Lee (1962) as:

$$Vf = \cos^2\ (90-H) \tag{5.10}$$

where Vf is the view factor (0.0 - 1.0); H is the average horizon angle from the zenith (degrees).

Incorporating the view factor into equation 5.9:

$$L_o{\downarrow} = (\varepsilon_a\ \sigma\ T_a^{\ 4})\ Vf + (\varepsilon_s\ \sigma\ T_s^{\ 4})(1 - Vf) \qquad (5.11)$$

This equation is only applicable for clear sky conditions and must be modified for use under varying types and amounts of cloud. Sellers (1965) offers the following equation to account for the amount, elevation and temperature of various cloud types:

$$L{\downarrow} = L_o{\downarrow}\ (1 + \Omega\phi^2) \qquad (5.12)$$

where Ω is a coefficient accounting for cloud height and temperature (appendix H); ϕ is the fraction of sky covered by cloud.

The complete longwave radiation balance over a snowcover may then be expressed as:

$$L^* = 1.24\ (e'_a\ /T'_a\ )^{.14}\ (P_a/1013)\ \sigma\ T_a^{\ 4}(1 + \Omega\phi^2)Vf + \varepsilon_s\ \sigma\ T_s^{\ 4}\ (1 - Vf) - \varepsilon_s\ \sigma\ T_s^{\ 4} \qquad (5.13)$$

In general, net longwave radiation will, in the presence of a snow covered surface, increase with elevation because atmospheric radiation decreases more rapidly with elevation than terrestial radiation (Kuzmin 1961).

5.3.1.4 Relative Contribution of Radiation Fluxes to Net Radiation

The importance of the individual radiation fluxes is known to vary on daily and seasonal cycles, and because of cloud cover, elevation, topography and overlying cover such as provided by forest canopies. In general, under clear sky conditions, it is direct beam shortwave radiation which is the most important for snowmelt production. Diffuse radiation, under these conditions contributes only 10% of the total global radiation (Male 1980), but for cloudy skies, it is more important than direct beam. Based on longterm radiation totals, Kondratyev (1969) has found that for the spring snowmelt season between 40^{o} and 60^{o} latitude, diffuse radiation accounts for 36 to 51% of global radiation.

Under clear skies there is usually a negative longwave radiation balance over the snowcover, largely because of the greater rate at which longwave radiation is emitted from the snow surface than from the water vapour in the atmosphere. During the daylight hours, this negative balance serves to reduce the positive net radiation balance produced by global radiation. At night in the absence of shortwave radiation, the net radiation balance is totally controlled by the effective radiation and becomes negative. Decreases in surface energy from a nighttime negative radiation balance can result in the lowering of surface snow temperatures and in the case of a melting snowpack, refreezing of the surface.

Under cloudy skies the situation is much different.

Dense clouds can also be considered to be black bodies and when the cloud temperature exceeds that of the surface snowpack, a positive effective radiation balance can develop. This may occur both in the day or nighttime. It might be expected that the contributions from K^* would exceed those of L^* during the daylight hours, but Ambach (1974) reported that when snow has a very high albedo, the effective radiation may be greater than the global radiation received at the snow surface.

The greatest net radiation totals are known to occur during partly cloudy conditions (Ambach 1974). In such cases, large contributions of global radiation are still received through gaps in the clouds and considerable amounts of longwave radiation are emitted by the cloud base.

5.3.2 Sensible and Latent Heat Flow

The terms Q_H and Q_E refer to the fluxes of sensible and latent heat to the surface of a snowpack. The magnitude and rate of these fluxes are dependent on the windstream, and the gradients of moisture and temperature above the snow surface, primarily within the first two to three metres.

The flux of both heat and moisture is composed of a part due to the mean vertical flow of air and a part due to eddying motion (Priestley 1959). Over the long term and at low heights above the snow surface, the mean vertical component becomes small (Swinbank 1951) and the eddying motion becomes the most important component of the transfer. This eddy flux is commonly referred to as the turbulent flux and may be measured with suitable equipment such as developed by Cramer and Record (1953), Dyer (1961) and Swinbank (1951). Although some measurements have been made over snow using this method [Hicks and Martin (1972); McKay and Thurtel (1978)], a more common method is the aerodynamic approach which involves numerous measurements of wind velocity, and temperature and moisture gradients above the snow (Priestley 1959). However, the differences between wind speed and humidity are extremely small within the atmosphere directly above a snowpack (Williams 1961) and hence impose stringent requirements on instrumentation. Secondly, the difficult nature of alpine environments and their complexity of wind conditions (Obled and Harder 1979) suggest the use of equations which although being slightly less accurate, require far less data input.

Many of the equations employed in the calculation of sensible and latent heat flow are termed bulk transfer equations. They require measurements of temperature and vapour pressure at only one height in the lower atmosphere and one at the snow surface. A logarithmic wind velocity profile is assumed and hence only a single measurement of wind speed may also be used. In general, the two transfer equations appear in the form:

$$Q_H = \rho_a \cdot C_a \cdot D_H \left(\frac{\Delta T}{\Delta z}\right) \qquad (5.14)$$

$$Q_E = \rho_a \cdot L_v \cdot D_E \left(\frac{\Delta q}{\Delta z}\right) \qquad (5.15)$$

where ρ_a is the density of air (kg m^{-3}); C_a is the specific heat of air (J kg^{-1} K^{-1}); L_v is the latent heat of vapourisation of water (J kg^{-1}); $\frac{\Delta T}{\Delta z}$ is the gradient of temperature between the snow surface and some height z (K m^{-1}); $\frac{\Delta q}{\Delta z}$ is the gradient of moisture between the snow surface and some height z (kg kg^{-1} m^{-1}); D_H is the transfer coefficient of sensible heat; D_E is the transfer coefficient of latent heat.

D_H and D_E represent the integrated effect of molecular diffusivity, through the thin laminar boundary layer directly above the snowpack and, eddy diffusivity through the turbulent layer (Price, 1977). The value of the transfer coefficients may be calculated from a wind profile based on a procedure which has been described in detail by many researchers [Sellers (1965); Priestley (1959); Anderson (1976); Price (1977)].

The main assumption used in deriving the coefficients is that they are equal to the exchange coefficient of momentum, D_M, which may be expressed as:

$$D_M = \frac{k^2 \; u_z}{\left[\ln \left(\frac{z}{z_o}\right)\right]^2} \qquad (5.16)$$

where k is von Karman's constant; u_z is the wind velocity at height z (m s^{-1}); z is the measurement height (m); z_o is a surface roughness parameter (m).

In most practical applications, windspeed is not measured at the same height as temperature or moisture (humidity). As first suggested by Sverdrup (1936) the denominator in equation (5.16) may be expanded for two measurement heights such that:

$$\left[\text{In } \frac{z}{z_o}\right]^2 = (\text{In } \frac{z_a}{z_o}) \; (\text{In } \frac{z_b}{z_o}) \qquad (5.17)$$

where z_a is the wind speed measurement height (m);
z_b is the moisture on temperature measurement height (m).

von Karman's constant has generally been found to equal 0.40 (Businger 1973), although some work by Businger et al. (1971) and Tenneks (1968) has pointed to a slightly lower value of 0.35. The surface roughness parameter, z_o, is frequently not measured but assumed to be equal to values which have been derived for specific surfaces. Lettau (1969) offers one method by which it may be calculated. Over relatively smooth snow, values of z_o have been found to vary between 5.0×10^{-3} m for a snow drifted wooded area (Price 1977) to as low as 5.0×10^{-5} for smooth snow over grass (Priestley 1959). When direct measurements of z_o have not been made, the value which is in common use is 2.5×10^{-3}, as suggested by Sverdrup (1936).

For the calculation of the moisture gradient, $\frac{\Delta q}{\Delta z}$, in equation 5.15, Rose (1966) has suggested the following equation for use within normal ranges of atmospheric humidity:

$$q = \frac{\gamma e_a}{P_a} \tag{5.18}$$

where q is the specific moisture of air ($kg\ kg^{-1}$); e_a is the vapour pressure of air (mb); γ is the ratio of molecular weights between air and water vapour; P_a is the barometric pressure mb.

The vapour pressure at the snow surface, e_s, may also be approximated by first assuming the air in direct contact with the ice matrix is saturated. Values of saturation vapour pressure over ice for varying temperatures may then be obtained from standard meteorological tables (List 1966). Hence, the moisture gradient may be expressed in vapour pressures such that

$$\frac{\Delta q}{\Delta z} = e_a - e_s \tag{5.19}$$

The temperature gradient in equation 5.14 may be calculated by measuring the air temperature T_a(K) at height z (m) and the surface snowpack temperature T_s(K), such that

$$\frac{dT}{dz} = (T_a - T_s) \tag{5.20}$$

Incorporating equations 5.16-5.17, 5.19 and 5.20 in equation 5.14 and 5.15, it is possible to write the equations for the transfer of sensible and latent heat as:

$$Q_H = \frac{\rho_a \cdot C_a \cdot k^2 \cdot u_z \, (T_a - T_s)}{\text{In} \, (\frac{z_a}{z_o}) \quad \text{In} \, (\frac{z_b}{z_o})} \tag{5.21}$$

$$Q_E = \frac{\rho_a \cdot L_v \cdot k^2 \cdot u_z \, (e_a - e_s) \gamma}{P_a \, \text{In} \, (\frac{z_a}{z_o}) \quad \text{In} \, (\frac{z_b}{z_o})} \tag{5.22}$$

5.3.2.1 Stability Corrections

The equations outlined above for sensible and latent heat assume a neutral atmosphere over the snowpack. However, during lapse (unstable) and inversion (stable) conditions, the transfer coefficients D_H and D_E do not necessarily equal that for momentum.

Stable conditions, whereby the surface snow temperature is less than that of the overlying air, is the most common situation during snowmelt. During conditions of low wind speeds the air directly above the snow is appreciably cooled and then being denser than the air above, will tend to remain in place. Turbulence is hence damped and the fluxes of sensible and latent heat reduced.

In order to account for such reductions, the bulk transfer coefficients may be modified by a stability factor. The simplest factor of this type is the well known bulk

Richardson number Ri (Richardson, 1920), which is defined as:

$$Ri = \frac{g \; z \; \Delta T}{T_a \; (\Delta u)^2} \tag{5.23}$$

where g is the acceleration due to gravity; T_a is the air temperature (K); ΔT is the temperature difference (K) over height z; Δu is the windspeed difference over the height z (m sec^{-1}); z is the height difference (m).

Monteith (1957) has used Ri in the following formula to correct the sensible heat transfer coefficient for stable conditions.

$$(D_H)_s = D_H/(1 + \alpha \; Ri) \tag{5.24}$$

where $(D_H)_s$ is the transfer coefficient for sensible heat under stable conditions; α is an empirical constant found by Webb (1970) to be approximately equal to 10.0.

Although Granger and Male (1978) report large decreases in the transfer coefficient when Ri was greater than 0.1, the greatest variation in sensible heat transfer from the neutral condition occurs in the unstable range [Granger (1977); Dyer and Hicks (1970); Swinbank (1968)]. A correction for unstable conditions similar to that for stable conditions is offered by Price and Dunne (1976).

$$(D_H)_u = D_H\ (1 - \alpha\ Ri) \tag{5.25}$$

where $(D_H)_u$ is the transfer coefficient of sensible heat for unstable conditions.

The effect of stability variations on latent heat transfer are relatively unknown, largely because of the difficulties investigators have found in accurately measuring humidity profiles. Even the few findings which have been reported show large discrepancies (Male 1980). This may indicate as noted by McBean and Miyake (1972) that for a passive scalar such as moisture (one which does not directly affect the buoyancy of the air), it may not be possible to establish a universal transfer relationship. In an attempt to at least partially account for this lack of information, Dunne and Leopold (1978) suggest the exact same adjustments to D_E as for D_H, hence

$$(D_E)_s = D_E/(1 + \alpha\ Ri) \tag{5.26}$$

$$(D_E)_u = D_E\ (1 - \alpha\ Ri) \tag{5.27}$$

where $(D_E)_s$ is the transfer coefficient for latent heat under stable conditions; $(D_E)_u$ is the transfer coefficient for latent heat under unstable conditions.

5.3.3 Precipitation Heat Flow

All forms of precipitation with a temperature greater than that of the snow on which it falls will transfer heat to the snow. In the case of an isothermal melting snowpack, only rain at temperatures greater than $0^{o}C$ is a potential heat source. Snow, sleet or super-cooled rain can act only as a heat sink.

The amount of heat which may be transferred from rain to a ripe snowpack is a function of the rain volume and temperature. The precipitation heat flow for rain may be expressed as:

$$Q_P = \rho_w C_w P (T_p - T_s) \times 10^{-3} \quad (5.28)$$

where ρ_w is the density of water (kg m^{-3}); C_w is the specific heat of water (J kg^{-1} K^{-1}); P is the precipitation intensity (mm s^{-1}); T_p is the temperature of the precipitation (K); T_s is the temperature of the snow surface (K).

If a heat deficit remains within the snowpack a second and much larger heat flow may be generated by rainfall. Rain in the presence of a large enough heat deficit will freeze and release latent heat to the surrounding snow. This heat flow may be written as:

$$Q_P' = P \rho_w L_f \times 10^{-3} \quad (5.29)$$

where Q_P' = precipitation heat flow due to latent heat only;

L_f = latent heat of fusion ($J\ kg^{-1}$);

and when combined with equation 5.28 the total precipitation heat flow becomes

$$Q_p^{\bullet} = P \rho_w \left[C_w (T_p - T_s) + L_f\right] \times 10^{-3} \qquad (5.30)$$

where $Q_p^{\bullet}$ is equal to Q_p' plus Q_p.

5.3.4 Ground Heat Flow

During the winter period there is normally a temperature gradient within the material underlying a snowcover. As the temperatures are warmer at depth there is a transfer of heat, primarily by conduction, upwards to the snowcover.

Calculation of ground heat conduction may be approached in two ways. The first employs a heat flux plate of known conductivity placed horizontally within the ground material. The temperature difference between the upper and lower faces of the plate is measured by a thermopile. The output is proportional to the temperature gradient across, and heat flux through the plate (Oke 1978) which can be considered approximately equal to the ground heat flow.

The second approach is used when the ground conductivity and temperature gradient are known. The heat flow may then be calculated by:

$$Q_G = \lambda \frac{dT_g}{dz} \qquad (5.31)$$

where λ is the ground thermal conductivity (W m^{-1} K^{-1}); $\frac{dT_g}{dz}$ is the temperature gradient within the soil (K m^{-1}).

Unfortunately, soil conductivities vary widely with porosity, mineral and organic content and moisture levels. Values of λ for various rock and dry soil types are presented by De Vries (1963) and Jumikis (1977), the latter also describing changes in conductivities with varying degrees of water saturation. In general, the overall conductivity of a ground material increases as the pore space is replaced with moisture.

The conductivity of ground materials are also known to increase with decreasing temperatures. The rate of increase for dry soil for each degree below freezing has been found by Penner (1970) to be approximately 0.1%, but lowering the temperature of saturated soil increases the conductivity much more rapidly. As the ground temperature decreases below freezing, the liquid water is increasingly converted to ice, which has a thermal conductivity four times that of liquid water.

A linear temperature gradient exists in the ground layers if the thermal conductivity is constant. However, beneath a snowcover the upper ground surface frequently has a higher conductivity than the underlying layers because of higher levels of moisture and ice content. If measurements of $\frac{dT_g}{dz}$ are made across such varying conditions, different values of λ must be used in equation 5.31.

5.4 Relative Importance of Heat Flows

The purpose of this section is to provide a general overview of the relative importance of the heat flows in equation 5.1 to a melting snowcover. More detailed discussions are introduced where appropriate in chapter six.

The significance of daily heat flows might be expected to vary according to elevation, latitude and time of year. However, Paterson (1969) noted, in a review of thirty-two energy balance studies from a variety of locations, that no consistent trends were apparent in the relative importance of the heat flows according to geographical location. This may partially be explained by the short term nature of many of the investigations. For instance, twenty-four of the thirty-two studies reviewed by Paterson (1969) present information concerning the length of measurement period and the average of these is only twenty-two days. The results from such short term periods would be highly dependent on the weather conditions prevailing at the time of study and would unlikely reflect long term trends. It is the long term results which would be expected to have a relationship with geographical location, but as noted by Colbeck et al. (1979) there is not sufficient data to undertake quantitative comparisons of the magnitude of the various energy terms according to climatic and physiographic factors.

Difficulties exist even in comparing short term results because of the variety of ways in which researchers present their findings. Percentages used to describe the heat fluxes do not always refer to a percentage of the total heat supply to the snowcover. For example, the heat consumed in evaporation is frequently subtracted from the heat supplied by radiation or condensation. In such cases, the real importance of the individual heat flows is masked. Another difficulty arises when comparing relative percentages from work conducted over melting snow versus snowcover which still has a large heat deficit. In cold snow, unlike for melting snow, the snow heat flow and storage may be considerable and hence reduce the relative importance of other terms. As mentioned, the following discussion deals only with results from ablation periods.

Despite the above mentioned difficulties, a number of useful comparisons can be drawn about the importance of the heat flows under a number of conditions.

Generally, net radiation and sensible heat are the major heat flows to a snowcover. Under clear skies and calm winds, especially in the spring and summer, net radiation normally dominates snowmelt production. Working under such conditions on Mt Blanc, de la Casiniere (1974) reported that net radiation contributed approximately 100% of the total heat supply to the snowpack (Q_H and Q_E each represented about 59% of the total heat exchange but were considered to balance each other). A similar result of 100% for Q^* was obtained by Lister (1962) for a much higher latitude location in the Canadian Arctic.

The greatest contributions of sensible heat occur under conditions of high temperature and windspeed. The relative importance of Q_H may be further enhanced if net radiation is small due to cloud cover or because of the sun angle as in polar regions. A useful contrast for high latitude areas can be made between the results of Lister (1962), described above, and Eriksson (1942). Although working at a comparable latitude, but slightly lower elevation than Lister (1962), Eriksson (1942) found net radiation contributed only 8% of the total heat supply with the majority (83%) coming from sensible heat. The differences in results are obviously related to prevailing weather locations and similar contrasts exist for other geographical locations. For many cases, as noted by McKay and Thurtell (1978), net radiation is usually the controlling parameter when air masses are well established while Q_H frequently becomes the most important when warm air is advected into a region. In terms of actual melt production, Granger and Male (1978) also point out that even when radiation is dominant, it is the sensible heat flow which may assist or counteract melt.

Sensible heat transfer has been found to be most consistently important in areas of low elevation. Fohn (1973) claims that at lower altitudes in mid-summer, Q_H is frequently equal to or greater than Q*. Within the European USSR Kuzmin (1961) has also found that sensible heat contributes 60 - 70% of the total heat to snow below 3000 m but only 40 - 50% above this elevation. In a review of energy balance results, Paterson (1969) also concluded that whenever Q_H was the dominant heat source, it was usually at the lower elevations. This pattern in the most part relates to the decrease in air temperature and increase in L* and K↓ with elevation (as noted in sections 5.3.1.1 and 5.3.1.3).

Condensation is normally found to rank third in importance of total heat supply during melt periods. Each volume of condensate added to the surface of a snowpack releases approximately 7.5 times the energy required to melt an equivalent volume of ice at $0^{o}C$. Large amounts of condensate can only be generated if the water vapour content of the overlying air is high. At low air temperatures, even if the relative humidity is high, the actual amount of moisture in the air is small. Hence, high air temperatures and humidities in conjunction with strong windspeeds are essential for a large condensation heat flow. In one particular study by Brazel and Marcus (1979) over a cold snowpack, Q_E was found to contribute 88% of the total heat supply even in the presence of cold, calm air. However, over melting snow the percentage is usually much smaller. Based on Paterson's (1969) review, Q_E ranged from 0 to a maximum of 30% of the total heat supply.

Precipitation heat flow, Q_P, is normally considered to be negligible during light rain events. Only when the amount, intensity and temperature of rainfall are high can Q_P transfer significant quantities of heat to a snowpack. The major ablation of snowcover has been frequently noted [for example: Price (1975); Colbeck (1975c); Bruce and

Clark (1966); U.S. Army (1956)] to occur during rain on snow events, which are characterized by reduced inputs of solar radiation, low cloud and heavy rain. For North America, these conditions have been found by Dunne and Leopold (1978) to be typical of the flow of strong moist air in the warm sector of cyclonic disturbances. In such situations, the major heat transfers are likely to be from Q_H and Q_E. Inputs of net radiation are strongly dependent on the size of the reduction in K↓ by cloud cover versus the increase in L↓.

On a daily basis the heat flow from conduction at the base of a snowpack is usually small and according to the U.S. Army (1956) produces only 0.5 mm d^{-1} of meltwater even under a moderately large soil-snow temperature gradient. Although, during snowmelt conditions this amount may be negligible, over an entire winter season the total accumulated heat flow can be important to the ripening of the snowpack and in the reduction of the total heat deficit.

Further comparisons of the relative importance of all heat flows within the New Zealand environment are described in the next chapter.

CHAPTER VI

SNOWMELT

6.1 INTRODUCTION

The main objective of this chapter is to investigate the snowmelt regime in the Craigieburn Mountains with particular reference to the importance of the various heat flows. Nine major sections are presented. The first reviews snowmelt-energy balance work which has been conducted in New Zealand and indicates areas of future research. The second section presents the results of field work carried out during the winter of 1980 to assess the heat deficit, over time, of the Craigieburn snowpack.

The next section discusses the methods of selecting snowmelt periods from the 1975-1980 meteorological record for energy balance calculations. This is followed by a review of the equations employed from chapter five and the methodology used in calculating the major heat flows. A separate section is also devoted to assessing the approximate value for ground heat flow, which is subsequently used as a constant in the discussion of snowmelt.

The remaining four sections detail the results of the heat flow investigations during selected spring and winter melt periods and consider the importance of days with high sensible heat transfer and the role of evaporation.

6.2 ENERGY BALANCE RESEARCH IN NEW ZEALAND

Some attention has been paid to the significance of the New Zealand snowpack as a water resource [Anderton (1973, 1974); Fitzharris (1972, 1979); Harrison (1978)] but little has been published concerning the details of snowmelt. A majority of the energy balance research has been conducted on glaciers, specifically the Ivory Glacier [Anderton (1976a, b); Anderton and Chinn (1978); Dickson (1974); Harding (1972)]. Results from these works provide some insight into the relative importance of the various heat flows to snowmelt in the New Zealand environment.

From a 34 day period in the summer of 1972 on the Ivory Glacier, Harding (1972) reported that over a snow and ice surface, radiation accounted for 40.4% of the total heat input, sensible heat 37.1%, latent heat 21.4% and heat from rainfall 1.1% [percentages have been corrected from the original results presented by Harding (1972)]. For the period in which the glacier was only snow covered, slightly higher percentages for sensible and latent heat were recorded than when the surface was snow free. As would be expected radiation values were lower over ice than snow, at least in part due to the reduction noted in the albedo of the surface.

Anderton and Chinn (1978) also present energy balance results for the same summer period, and although the analysis period was reduced to 30 days, their results differ considerably from those of Harding (1972). In particular, radiation was calculated to have accounted for 53% of the total heat supply, latent heat only 11% and sensible heat was noted at a similar value of 33%. They also present results from a seventeen day period in the summer of 1973, during which the percentages were 51, 25, 22 and 2 for the radiation, sensible, latent and precipitation heat flows respectively. This again points to the dominance of radiation but as noted by Anderton and Chinn (1978) these results are characteristic only of relatively calm periods dominated by clear skies.

It should be noted that the Harding (1972) results were corrected for minor errors and to account for Q_E (latent heat of condensation) only as a heat input. This latter adjustment was not possible for the Anderton and Chinn (1978) results as they presented only a summary table and no indication whether evaporation (negative value of Q_E) was subtracted from the daily heat totals. Adjustments to the Harding (1972) data produced only minor changes because of the lengthy study period and the minor role of evaporation. The same is expected for the Anderton and Chinn (1978) results.

It has generally been regarded that the seasonal snowcover in New Zealand has the highest ablation rates during rain events which may occur even in the winter season [Fitzharris (1979); O'Loughlin (1969a); Prowse (1980)]. Anderton and Chinn (1978) reported a two day rain event which produced 354 mm of runoff (errata: 454 mm was referenced but the original report by Anderton (1976b) quotes 216 mm plus 138 mm on two days of rainfall). The precipitation heat flow from this event accounted for 19% and 11% of the daily heat inputs on two subsequent days. They also noted that for one other storm, in November 1974, the heat content of rain was sufficient to produce 45% of the recorded ablation during the event. Fitzharris et al. (1981), reporting on the contribution of snowmelt to a spring flood in Otago, also points to the importance of the precipitation heat flow. Based on observations of water content loss in an alpine snowpack and the extrapolation of low elevation meteorological data, he calculated that rain contributed 23% of the total heat input over a 43 hour period. This exceeded the total supplied by radiation at 20% but was less than the remaining percentage supplied by sensible and latent heat transfers.

As the above brief summary demonstrates, the amount of information concerning heat flows to a melting snowcover is scarce and of variable quality. Although some energy balance work has been conducted, it has been primarily over ice surfaces and during calm, clear sky, summer periods. In addition, the only results dealing with a seasonal snowcover have not involved on site measurements, but rather the elevational extrapolations, of heat flows. Surprisingly, no results have been presented for winter melt periods, which have been reported to be so significant in the New Zealand maritime environment (Fitzharris 1979).

Because of the scarcity of snow pit data, no information exists concerning the heat deficit of the seasonal snowcover. This type of information can be very useful in assessing the melt-responsiveness of a snowpack over time. Some rates of snowmelt runoff have been reported by O'Loughlin (1969b) and Fitzharris (1979) but comparative rates do not exist for varying weather and snowpack conditions.

6.3 HEAT DEFICIT

As described earlier, the heat deficit of a snowpack is a measure of its resistance to runoff production. The greater the mass and the lower the mean temperature of a snowpack, the greater the heat deficit. A one metre polar snowpack with an average density of 350 kg m^{-3} and mean temperature of $-20^{\circ}C$ would have the same heat deficit (14.6 MJ) as an alpine snowpack of similar density but with a mean temperature of only $-5^{\circ}C$ and four metres in depth. If both snowpacks were of equal depth, the polar snow would require approximately four times the energy input as the alpine snowpack before becoming isothermal at $0^{\circ}C$ and hence be capable of producing runoff.

Although a snowpack usually has a zero or near zero heat deficit in the spring months, it will usually be at a maximum during the winter months. Knowledge of the winter heat deficit can allow an accurate assessment of the amount of energy which is required to produce meltwater runoff.

In order to establish the approximate heat deficit of the Craigieburn snowpack, the snow pit record from the 1800 m level on Mt Cockayne was analyzed. As the heat deficit increases with elevation, because of colder temperatures and greater snow accumulation, the following results should be considered to be representative of the high elevation heat deficit for the Craigieburn Mountains.

The heat deficit would also be expected to vary markedly between sunny-north and shady-south facing slopes. Hence, snowpit records were analyzed for both aspects.

The heat deficit for a given one cubic metre column of snow from a snowpack may be calculated from:

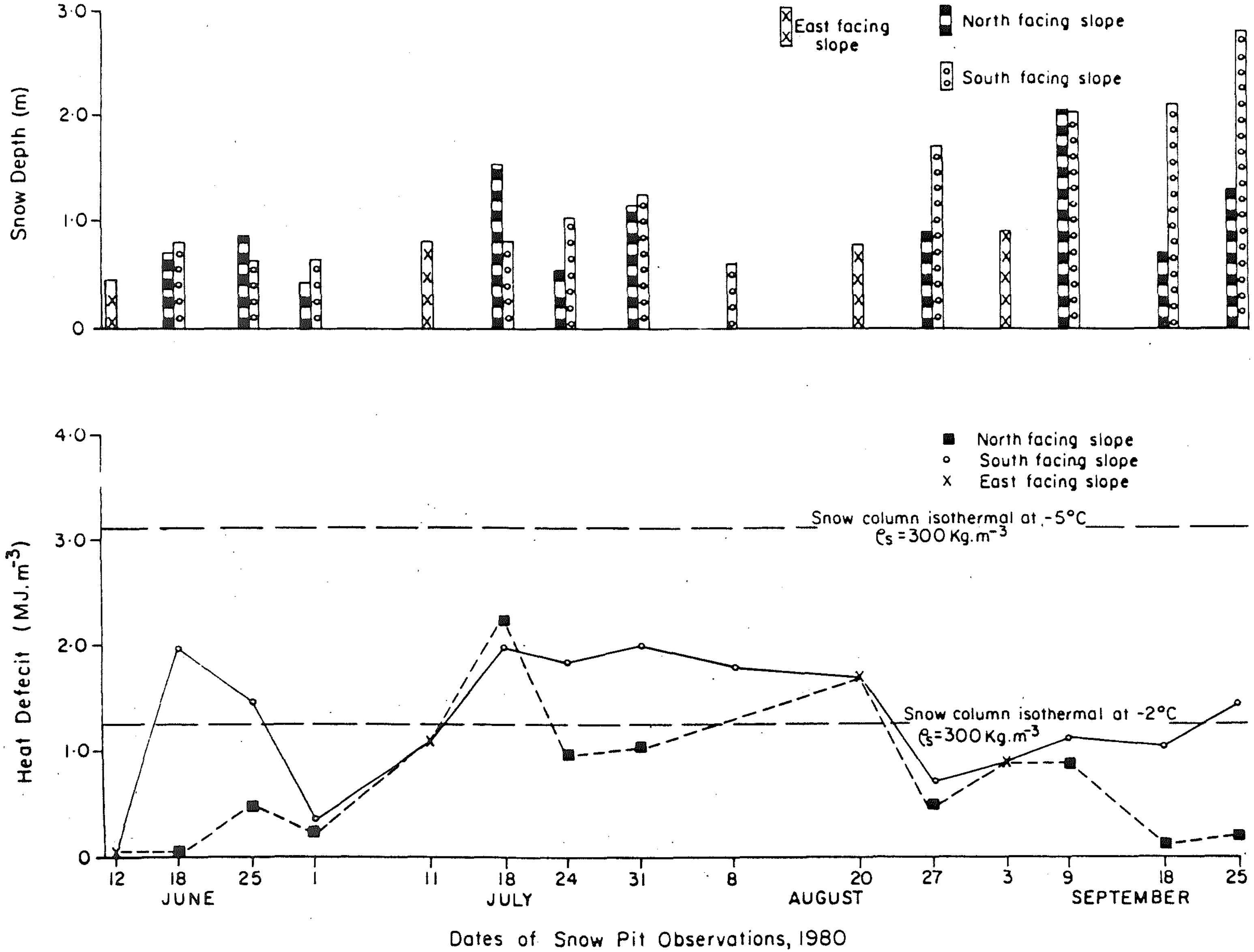
$$\zeta = \sum_{j=1}^{n} (z_j \cdot \rho_{s_j} \cdot T_{s_j} \cdot C_i)/d_s \qquad (6.1)$$

where ζ is the heat deficit (J m^{-3}); n is the number of different stratigraphic layers within the snowpack; z_j is the thickness of layer n (m); d_s is the total thickness of snowpack (m); ρ_s is the snow density of layer n (kg m^{-3}); T_{s_j} is the snow temperature of layer n (K); C_i is the specific heat of ice (J kg^{-1} K^{-1}).

The results of this analysis are presented in figure 6.1.

Although the concept of a heat deficit is referred to in much of the literature pertaining to snowmelt, it has not been used to classify the coldness of a snowpack. Much more simple methods have been employed such as that offered by Smith (1974), who classifies a snowpack as being warm or cold if the temperature at 300 mm below the surface is near or well below the melting point. Based on the 1980 snow pit record, the Craigieburn snowpack would have to be classed as relatively warm. For a majority of the season, the -300 mm temperature remained between -2 and -5^{o}C. Extreme snowpack temperatures were limited to only the surface layers which were observed to reach a daytime minimum

Figure 6.1 HEAT DEFICIT AND SNOW DEPTHS ON OPPOSING ASPECTS, MT COCKAYNE 1980. Heat deficit values were calculated from snow pit observations at an approximate elevation of 1800 m on primarily south-east and north-east facing slopes. The total heat deficit can be obtained by multiplying the depth (m) by the listed heat deficit values. At the end of September the snowpack had become isothermal at $0^{o}C$.


Snow Depth (m)
3·0
2·0
1·0
0
East facing slope
North facing slope
South facing slope
Heat Defecit (MJ.m^{-3})
4·0
3·0
2·0
1·0
0
North facing slope
South facing slope
East facing slope
Snow column isothermal at -5°C
ρ_s = 300 Kg.m^{-3}
Snow column isothermal at -2°C
ρ_s = 300 Kg.m^{-3}
12
18
25
1
11
18
24
31
8
20
27
3
9
18
25
JUNE
JULY
AUGUST
SEPTEMBER
Dates of Snow Pit Observations, 1980

temperature on July 18 at -12^{o}C.

Although strong fluctuations are apparent in the size of the heat deficit in figure 6.1, for most of the winter it remains small. Based on a representative one metre snow depth, only once during the 1980 season did the heat deficit exceed 2 MJ m^{-3}. This occurred on July 18 at which time the total heat deficit on the north-east facing slope was over 3.5 MJ. Because of lower snow depths on the south-east slope, the total heat deficit was about half at 1.7 MJ.

Interestingly, the largest total heat deficit of 4.1 MJ was recorded at the end of September on the south-east face. This large deficit may be attributed to the overall mass of snow which reached 2.8 m in depth, rather than cold temperatures. In fact, considerable melt of the north facing slope had already occurred.

In summary from the 1980 results, the snowpack of the Craigieburn Mountains, even at high elevations, may be classed as warm. Surface snow temperatures may reach well below -10^{o}C, but the total heat deficit remains small because of low mean temperatures. The size of the heat deficit is such that only one day of strong heat input distributed throughout the snowpack would be sufficient to eliminate the total heat deficit. The magnitude of these inputs is described in the following sections.

6.4 SELECTION OF MELT PERIODS

In order to evaluate the size and relative importance of the various energy components to snowmelt, specific melt periods had to be selected. The original intention was that such periods would be based directly on runoff records from Camp Stream basin. However, during the field winters of 1978, 1979 and 1980 insufficient snowfall precluded the use of the basin for this particular aspect of the study.

As an alternative, melt periods were selected from snow depth records made at the SB climate station. A number of obvious problems exist in the use of simple snow depth as an indication of snowmelt. Firstly, changes in depth may reflect the action of wind deflation and erosion. Secondly, decreases in snow depth can be due to the natural process of settlement from equi-temperature metamorphism. Finally, snow depth measurements alone do not provide any information concerning changes in the water equivalent of a snowpack. They therefore cannot be used in the estimation of snowmelt runoff without the added information provided from density measurements.

In view of these problems, specific methods were used in the selection of snowmelt periods. Only periods of rapid snow depth depletion, covering several days and preceded by a precipitation-free period of relatively constant snow depth, were included for final selection. This ensured that the depth decreases were not simply due to settlement following new snow inputs. Because temperatures in the Craigieburn Range are relatively warm, a majority of settlement from equi-temperature metamorphism will occur in the few days immediately following a snowstorm. Settlement rates would only be slight after this phase and rapid decreases in depth would more than likely be due to melt.

The periods selected by this method were characterized by near surface air temperatures well above the freezing mark. This in itself is indicative of a melt condition. Since the surface snowpack was likely to be in a state of melt, the surface shear strength would be relatively high and would reduce the possibility of wind erosion. The snow depth measurement location was also believed not to be a zone of noticeable wind deflation. In times of strong winds, as much snow was usually deposited as eroded from the site.

Based on the above selection techniques and on the availability of meteorological information, nine melt periods were finally selected. The following describes the methods by which the components of the energy balance were evaluated on a daily basis.

6.5 CALCULATION OF HEAT FLOWS

The following details the formulae and methods used in the calculation of the components. Many of the formulae have already been outlined in chapter five.

6.5.1 Radiation Heat Flow

Records of incoming shortwave radiation, $K\downarrow$, were obtained from the actinograph located at the CF climate station. Daily values were adjusted for the SB site based on the results of a six year comparison between the two sites (appendix I). As described in chapter five, K↓ should normally decrease at lower elevations because of increased back-scattering. However, during the main winter period, the SB site had lower values of K↓ than CF. This discrepancy may be explained by differences in view factors relative to the solar path, localized cloud or obstruction by snow, ice and riming. The relative importance of these factors is unknown and hence it was decided to adjust the daily K↓ values according to the results in appendix I. The implications of these adjustments are described in the results sections.

Reflected shortwave radiation, K↑, is directly dependent on the surface albedo. No direct measurements of albedo were made but as earlier described, there are relationships known to exist between surface snow properties and albedo. Anderson's (1976) formula establishes such a relationship between surface snow density and albedo. Based on one hundred and six separate surface snow density measurements made over the 1980 winter season at the SB site and using Anderson's (1976) formula, mean monthly albedo values were estimated (table 6.1).

Surface snow albedos are also generally known for melt conditions. Kuzmin (1961) reports the mean albedo of a clean melting snow as 0.72. Similar results of approximately 0.70 are presented by Bergen (1975); Dunne and Leopold (1978); Kuznetsov (1960) and U.S. Army (1956). These relatively high values refer to the early stages of melt especially during the accumulation season. In the later stages of melt, when the surface becomes highly

MONTH	MEAN SURFACE ρ_s (kg m-3)	SAMPLE SIZE	a_k
JUNE	224	17	0.84
JULY	232	29	0.81
AUGUST	275	23	0.76
SEPT	312	21	0.72
OCT	376	16	0.63

Table 6.1 SURFACE SNOW DENSITY AND CALCULATED ALBEDO
Mean surface snow densities, ρ_s, were measured during the 1980 winter season at Broken River meteorological station. Values of the albedo, a_k, are calculated according to the relationship established by Anderson (1976).

granular, the albedo falls to much lower values. Dunne and Leopold (1978) quote a range between 0.5 and 0.6, while the U.S. Army (1956) and Maguire (1975) quote minimum values of 0.4 and 0.25 respectively for old dirty snow surfaces.

Considering that the surface albedo values in table 6.1 were derived from both melt and non-melt periods, and in view of the results from other investigations, a value of 0.7 was employed for brief melt periods during the winter season and 0.5 for the late season spring melt.

Values for net longwave radiation, L*, over the snowpack were calculated using equation 5.13. Temperature, vapour pressure and atmospheric pressure values were extracted in the same manner as for sensible and latent heat, which is outlined in section 6.5.2. Information detailing the cloud height, amount and type was obtained from the daily climate records at CF. Based on equation 5.10 and eight (45^{o} separation) compass bearings, the view factor for the SB site was calculated to be 0.958.

6.5.2 Sensible and Latent Heat Flows

Transfers of sensible and latent heat to the snowpack were based on equations 5.21 and 5.22, and adjusted for stability as described in section 5.3.2.1. Temperatures used in these equations and that for longwave radiation (equation 5.13) were extracted from thermograph records eight times during the day beginning at 06:00 and every three hours afterwards. All chart values were adjusted according to daily time and temperature checks. Similarly, relative humidity values were extracted from charts but these had to be converted to vapour pressure for the SB site according to the psychrometric equation:

$$e_a = e_{\nu w} - \left(\frac{C_a}{\gamma}\frac{P_a}{L_v}\right) \quad T_d - T_w \qquad (6.2)$$

where e_a is the vapour pressure (mb); $e_{\nu w}$ is the saturation vapour pressure (mb) at the wet bulb temperature; C_a is the specific heat of air (J kg^{-1} K^{-1}); P_a is the near surface air pressure (mb); γ is the molecular weight ratio of air to water vapour (0.622); L_v is the latent heat of vapourization (J kg^{-1}).

Measures of windspeed were obtained from a totalizing cup anemometer located at the SB site. Windspeeds had to be averaged between the times of morning observations, commonly between 10:00 and 11:00. As a result, the average windspeeds do not directly coincide with the daily time frames used for the other variables. The magnitude of the overlap is small and because of the consistency of windspeeds, encountered during melt periods, only minor inaccuracies were expected.

Standard instrument heights, z_a and z_b, at the SB site are 2.7 m for the wind recorder and 1.4 m for the temperature and humidity measurements. However, for equations 5.21 and 5.22 these heights are applicable only in the absence of snowcover. The effective instrument heights had to be reduced for the depth of snowcover which changes over the season. Instrument heights were calculated by:

$$z_a' = z_a - d_s \qquad (6.3)$$

$$z_b' = z_b - d_s$$

where z_a' is the adjusted height of the wind instrument (m); z_b' is the adjusted height of the vapour pressure or temperature measurement (m); d_s is the depth of snowcover (m).

6.5.3 Precipitation Heat Flow

Precipitation heat flow was based on precipitation gauge and temperature records, and equation 5.28. Because the temperature of the rain was not measured directly, wet bulb air temperatures were used as an approximation, as suggested by Anderson (1976). Wet bulb temperatures were calculated from the humidity and temperature records for each rain event. Numerous measurements were made for lengthy events. On a few occasions the wet bulb temperature was calculated to be sub-freezing and hence the heat flow was treated as zero. The addition of energy from the release of latent heat was included any time the snowpack was considered to have a heat deficit.

6.5.4 Ground Heat Flow

Direct measurements of the ground heat flow were not made at the SB site during the 1975-80 winters. The only consistent sub-surface temperature recording was made at 100 mm below the surface. As mentioned, ground heat flow is normally small in comparison to the other components of the daily energy balance. For many applications an approximation of the heat flow is made and used as a constant in daily melt computations.

In order to obtain an approximation of Q_G, equation 5.31 was used in conjunction with measurements of the ground conductivity and temperatures from the Mt Cockayne thermistor profile. The assumption was made that the ground heat flow through scree material, beneath a snowcover, at a comparable elevation on Mt Cockayne would be similar to those at the SB site.

6.5.5 Ground Heat Flow Results

The Mt Cockayne thermistor profile used in the assessment of ground heat flow operated during both the 1979 and 1980 winter seasons. Breaks in the record were mainly due to the failure of the system from lightning strikes.

6.5.5.1 Temperature Gradients

Degree hours were summed on a weekly basis for the -50 and -150 mm thermistors over the 1979 winter in a first attempt at obtaining a record of the sub-surface temperature gradient. Small differences in accumulated degree hours precluded the use of this method. The results appear in appendix J.

The original temperature record was then analyzed at fifteen minute intervals to obtain a more accurate estimate of the gradient. Based on results from this method and known snow depth coverage, two periods were selected for further study. They are considered to be representative of temperature conditions during periods of initial snow coverage (figure 6.2a) and established thick snow cover (figure 6.2b).

Prior to June 29 the ground at the number six thermistor site was snow free. As illustrated in figure 6.2a, up until this date both thermistors recorded sub-freezing temperatures and pronounced fluctuations existed in the -50 mm temperature record. This temperature pattern can be explained by the direct thermal contact existing

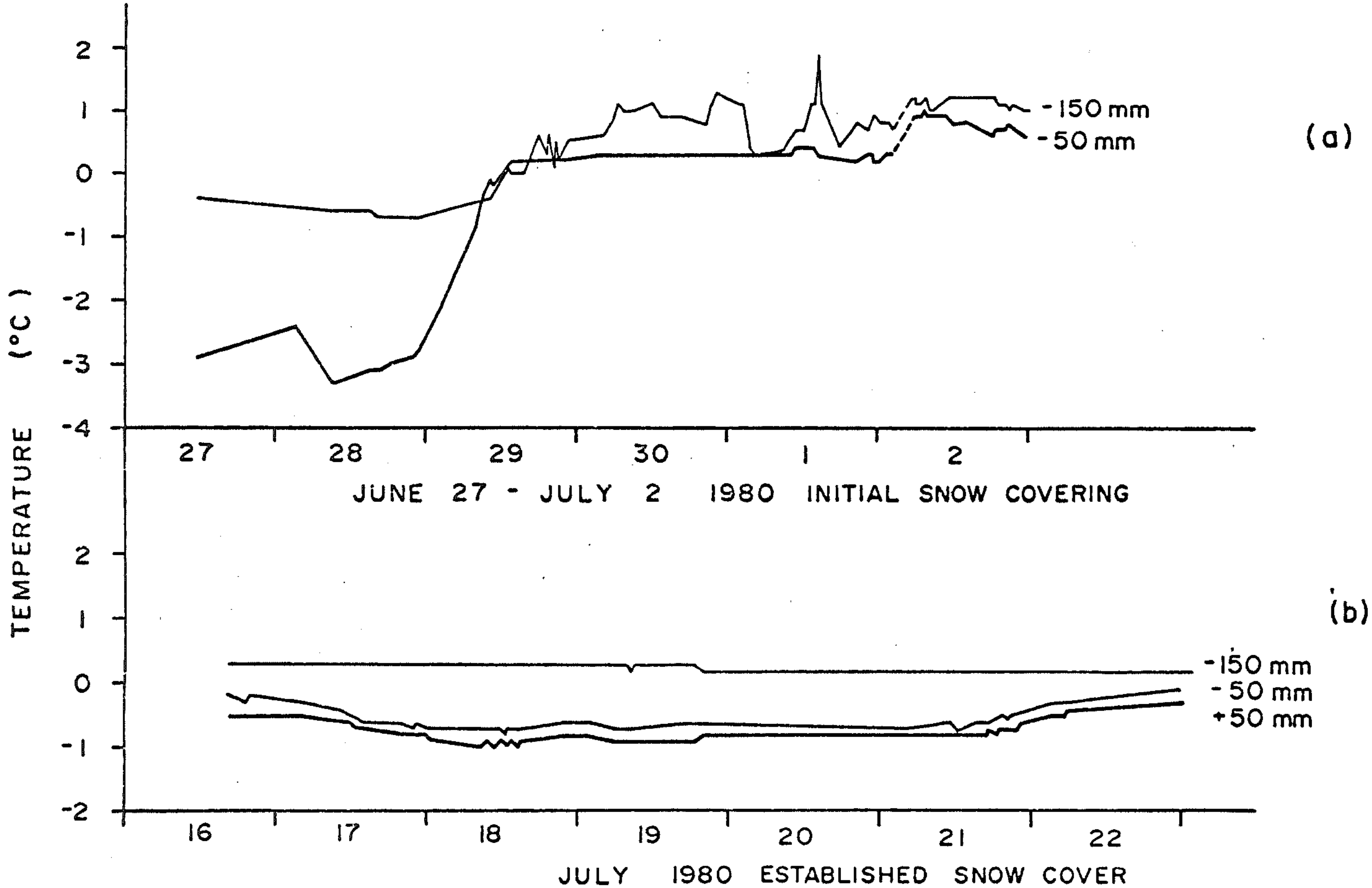


Figure 6.2a,b GROUND TEMPERATURES BENEATH SNOWCOVER. The upper diagram (a) illustrates temperatures 50 and 150 mm beneath the snow-ground interface at the time of initial snow covering. The lower diagram (b) represents the same ground temperatures plus one snow temperature (+50 mm) once the snowcover was fully established.

between the ground and the atmosphere, and because of unrestricted heating and cooling from radiative exchanges.

A thin snowcover was established on June 29 and effectively insulated the ground surface to all energy exchanges. This produced a damping of temperature fluctuations and a general increase in ground temperatures. The -50 mm thermistor recorded temperatures just near the freezing mark while the -150 mm was characterized by above freezing temperatures. During this phase the near surface temperature gradient ranged from approximately 1 to 5 K m^{-1}

Once a full depth of snowcover became established a stronger temperature gradient existed between the -150 and -50 mm thermistors. As illustrated in figure 6.2b the -150 mm thermistor remained relatively constant at just above freezing while the -50 mm thermistor record was dominated by slow long term fluctuations. The temperature gradient over the period ranged from approximately 3 to 10 K m^{-1}.

However, the gradient between the two ground thermistors may not be representative of the gradient existing between the ground surface and the snowpack. In the calculation of Q_G based on the results from figure 6.2b another problem also arises. During the July 16 - 22 period, one thermistor recorded sub-freezing temperatures, the other above freezing. As explained in chapter five, soil conductivities markedly vary between frozen and unfrozen soils and hence a linear temperature gradient cannot be assumed in equation 5.31.

Figure 6.2b also illustrates the +50 mm thermistor temperature trends which can be considered representative of temperatures near the base of the snowpack. The trend of temperatures from this thermistor closely mirrors that for the upper ground layer, the gradient between the two remaining relatively constant at 1 to 3 K m^{-1}. The actual gradient between the -50 mm thermistor and the ground-snow interface would be less than this but there is no practical means of further assessing this gradient. The ground-snow temperature gradient is therefore assumed to be approximately 1 - 3 K m^{-1} for full depth winter snowcover over scree covered surfaces.

6.5.5.2 Thermal Conductivities

The selection of appropriate thermal conductivity values, λ, for use in equation 5.31 depends very much on the composition of the ground material and whether it is in a frozen state. Greenland (1971) conducted some experiments on alpine soils in the Cass region and concluded that values for λ varied from approximately 1.39 to 2.37 W m^{-1} K^{-1}, with a single high value of 5.56 W m^{-1} K^{-1}. Greenland (1971) concluded the high value was due to the effect of freezing. This explanation seems unlikely in view of Jumikis' (1977) report that the highest λ for even sandy soil at 100% saturation and 42% ice content was only 2.96 W m^{-1} K^{-1}. Similarly, Penner (1970) noted that for a silt-clay soil at two degrees below freezing and partially frozen, the thermal conductivity was only 2.09 W m^{-1} K^{-1}.

The soils in the upper elevations of the Craigieburn Range are known to have a typically shallow horizon of silt-loam to fine sandy-silt loam. The soil forming parent materials, often exposed at the surface, consist of indurated sandstone and siltstone including loess, colluvium and scree (Wilde 1974).

In view of all of the above, the assumption is made that the upper frozen soil layers in the study area have an approximate range of thermal conductivities between 2.0 and 2.5 W m^{-1} K^{-1}.

However, the actual placement of the thermistors used for assessing the temperature gradients was in a zone of scree-soil. O'Loughlin (1965) has defined the composition of scree materials in the area to be 70 - 75% boulders and gravels 2 mm in diameter, 25 - 30% sand .02 - 2.0 mm, and 1.5 - 2% silt 0.2 mm. The boulders are primarily a form of sandstone for which Jumikis (1977) reports a range in λ from 1.0 to 2.5 W m^{-1} K^{-1}. The higher range is not dissimilar to the suggested thermal conductivities for the soil.

6.5.5.3 Results

Based on the calculated temperature gradient of 1 - 3 K m^{-1} and a thermal conductivity range of 2.0 to 2.5 W m^{-1} K^{-1}, values for ground heat flow were calculated from equation 5.31 to be in the range of 0.17 to 0.65 MJ d^{-1}. The results are similar to many others reported in overseas research. Yoshida (1962) in Japan quotes rates of 0.26 to 0.36 MJ d^{-1}, and in Canada, Granger (1977) and Gold (1957) report values of 0.00 to 0.26 and 0.86 MJ d^{-1} respectively. Kuzmin (1961) in the European U.S.S.R. places an approximate range of $\pm$0.40 MJ d^{-1} on ground heat flow below snowcover.

The calculated values for the Craigieburn Mountains refer only to open ground surfaces without vegetation. Although most of the upper elevations are free of vegetation, much of the middle elevations above treeline are covered in grasses particularly tussock. As noted by Wilson (1941), heat transfer can be significantly reduced by surface vegetation. The effect of this is unknown for the Craigieburn site. However, it was observed during snow pit excavation that soil under tussock plants remained unfrozen while in clear areas the surface was frozen.

The above discussion has focused on conditions when the snowpack is characterized by below freezing temperatures. Once the heat deficit has been eliminated from the snowpack (snow temperature isothermal at $0^{o}C$), changes in the ground heat flow may result if the ground material remains frozen. By comparing dates of snowmelt and ground temperatures, the direction of the heat flow may be deduced.

Figure 6.3 illustrates the long term mean ground temperatures recorded from the -100 mm sensor located at the SB meteorological station [after Child (1978)]. Although the mean values for the period June to October are just above freezing, for most years the soil was frozen at -100 mm from late June to mid September [Apse (1967-1969); Watson (1970-1974); Noble (1975); Child (1978-79)]. Because the snowcover at the same elevation normally becomes isothermal at $0^{o}C$ by at least October, for much of the spring season there would be a heat flow away from the snowpack.

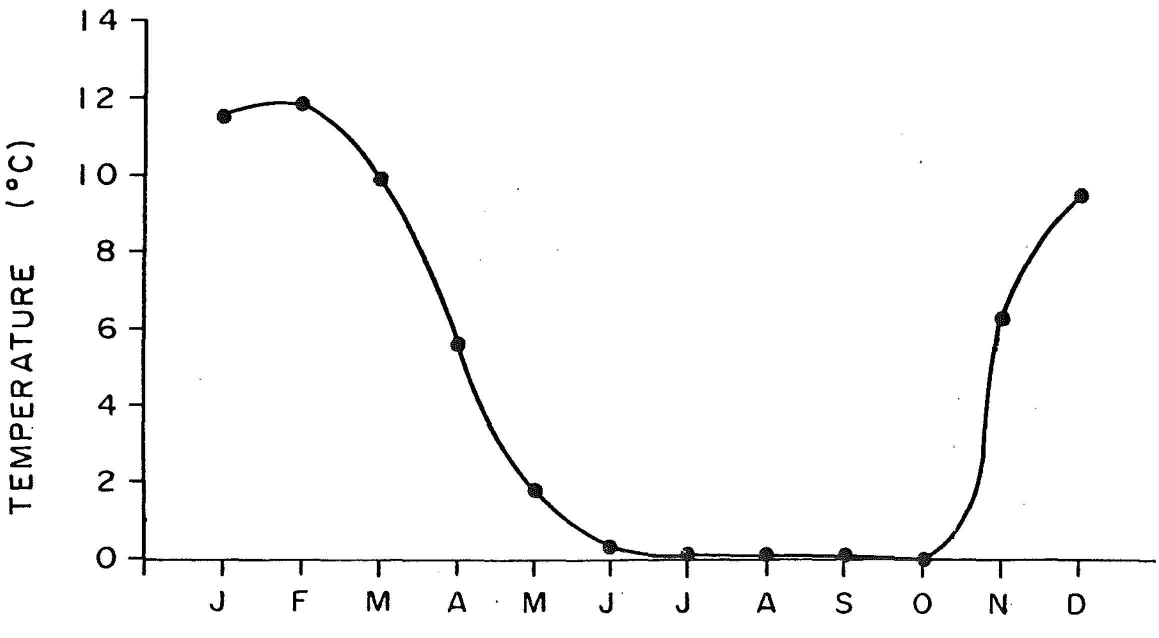


Figure 6.3 MEAN MONTHLY SOIL TEMPERATURES AT -100 mm. Monthly averages were derived from mean values over the period 1969-1978 at the SB climate station.

6.5.5.4 Conclusion

For a majority of the winter season ground heat flow was found to be relatively small, contributing no more than 0.65 MJ d^{-1}. This flux would be sufficient to melt 2 mm of water from the base of the snowpack.

During early spring there is likely a heat flow away from the snowpack to the frozen ground below. Q_G cannot therefore be considered as a heat source for the onset of snowmelt. The importance of ground heat flow once the ground is unfrozen is unclear, but daily inputs would not be expected to exceed those recorded for the winter period.

As ground heat flow is small relative to the other heat sources reported in subsequent sections, it will not be considered in the daily sums of that for snowmelt periods.

6.6 WINTER MELT PERIODS

6.6.1 General

According to the methods outlined in section 6.4 five ablation periods were selected from the 1975-80 meteorological record between the months of May and September. The final selection of periods was primarily controlled by the availability and accuracy of meteorological records from which daily energy balances had to be constructed.

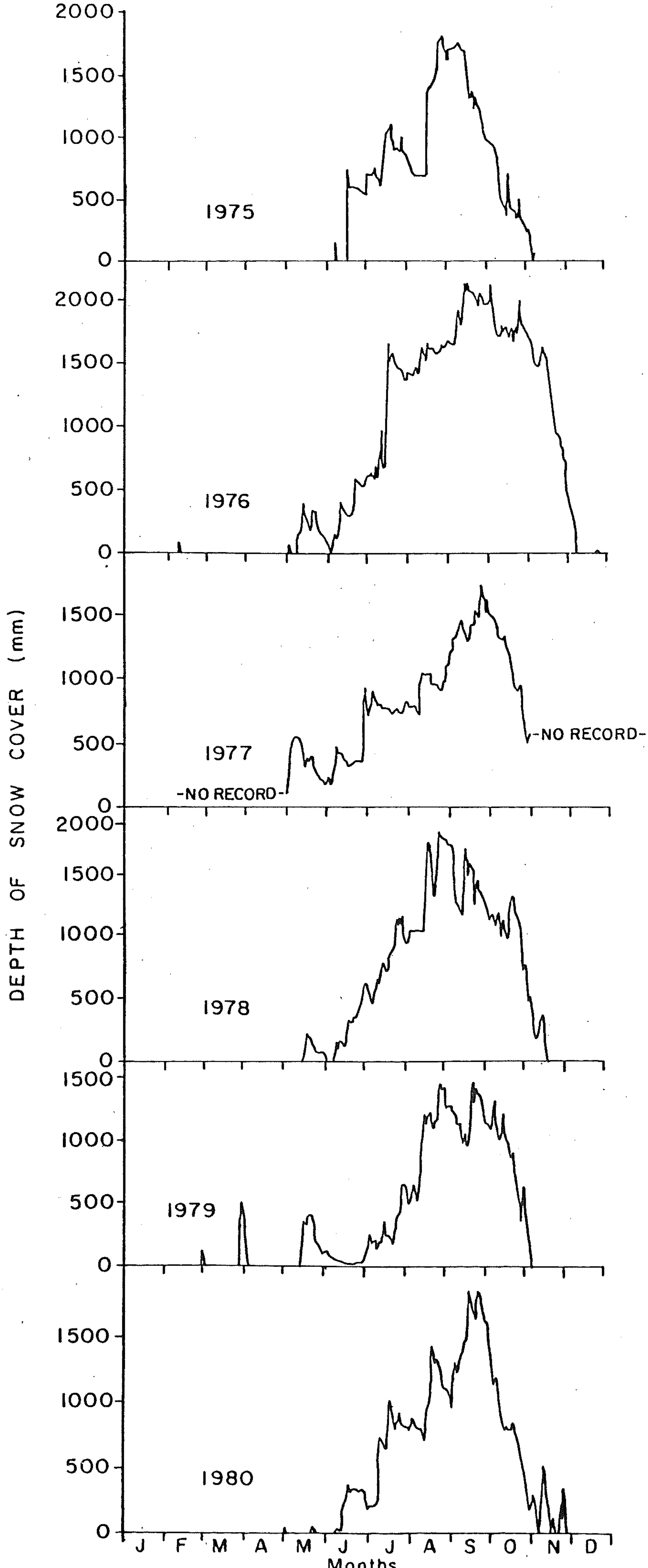
6.6.2 Description of Melt Periods

Three of the melt periods occurred during May, one during 1979 and the other two in 1977. These early season periods are characteristic of years (figure 6.4 and section 4.4.2.3) which receive heavy snowfalls in late April or May, but which also experience subsequent warming trends before the main winter season. During 1980 the main accumulation period did not begin until July, and was preceded by the fourth melt period, June 29 - July 2.

The fifth melt period occurred in mid-September 1979 and effectively bisected the snow accumulation curve for that year. Although this period occurred late in the year, after the lowest ebb of the solar radiation cycle, it is distinct from the late season melt periods discussed in the next section. The main distinction is that the spring melt periods occurred in months dominated by above freezing temperatures and significantly higher available solar radiation.

The general weather conditions which prevailed during the five winter melt periods are summarized in table 6.2. Importantly, all periods were characterized by windspeeds in excess of monthly averages and receipts of solar radiation below average. In addition all periods, except that of May 25 - 28, recorded above average air temperatures. This was also the only period which had relatively clear skies while others were dominated by extensive and frequently low cloud cover. The specific weather conditions on days of exceptional melt and energy inputs are discussed later.

Figure 6.4 SNOW DEPTHS AT THE SB CLIMATE STATION 1975-1980. The curves are based on daily snow depth measurements taken against a snow pole. Note most major increases occur during late July and the month of August.


DEPTH OF SNOW COVER (mm)
1975
1976
1977
-NO RECORD-
-NO RECORD-
1978
1979
1980
0
500
1000
1500
2000
J F M A M J J A S O N D
Months

	AIR TEMPERATURES ($^{\circ}C$)		MAXIMUM AIR TEMPERATURES (MORNING OBSERVATIONS) ($^{\circ}C$)		WINDSPEED ($m\ s^{-1}$)		GLOBAL RADIATION ($MJ\ d^{-1}$)		VAPOUR PRESSURE (mb)	
	MEAN	DIFF.	MEAN	DIFF.	MEAN	DIFF.	MEAN	DIFF.	MEAN	
MAY 12-15, 1977	4.0	+2.1	5.7	+0.9	4.2	+1.7	5.15	-0.02	5.36	
MAY 21-23, 1977	3.3	+1.4	5.6	+0.8	3.7	+1.2	3.69	-1.48	5.75	
MAY 25-28, 1979	1.7	-0.2	3.3	-1.5	4.0	+1.5	4.84	-0.33	5.38	
JUNE 29 - JULY 2, 1980	2.0	+2.9	3.3	+1.4	3.7	+0.9	4.12	-2.04	5.90	
SEPTEMBER 10-14, 1979	2.9	+2.8	4.8	0.0	4.4	+1.0	6.95	-2.04	6.14	

Table 6.2 WEATHER CONDITIONS DURING WINTER SNOWMELT PERIODS. MEAN values refer to those calculated over the respective study period. DIFF. refers to the difference between the MEAN value and the longterm mean monthly value. Note: for the period June 29 - July 2 the mean monthly value was averaged for June and July in the calculation of DIFF.

6.6.3 Heat Flow Results

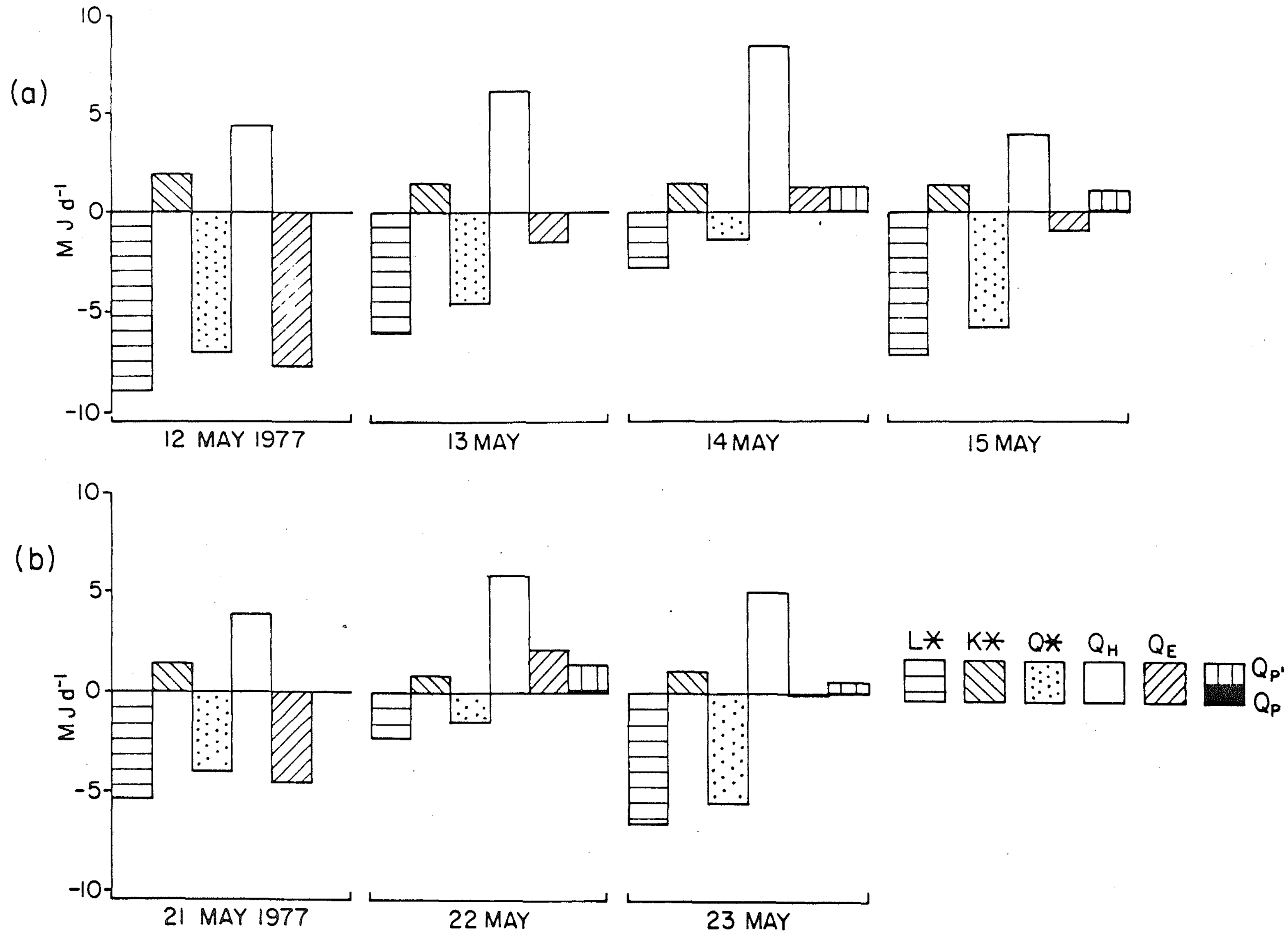
Based on the methods outlined in chapter five and section 6.5 of this chapter, the heat flows were calculated on a daily basis for all five melt periods. The results are presented in figures 6.5a to 6.5e and summarized in tables 6.3 and 6.4.

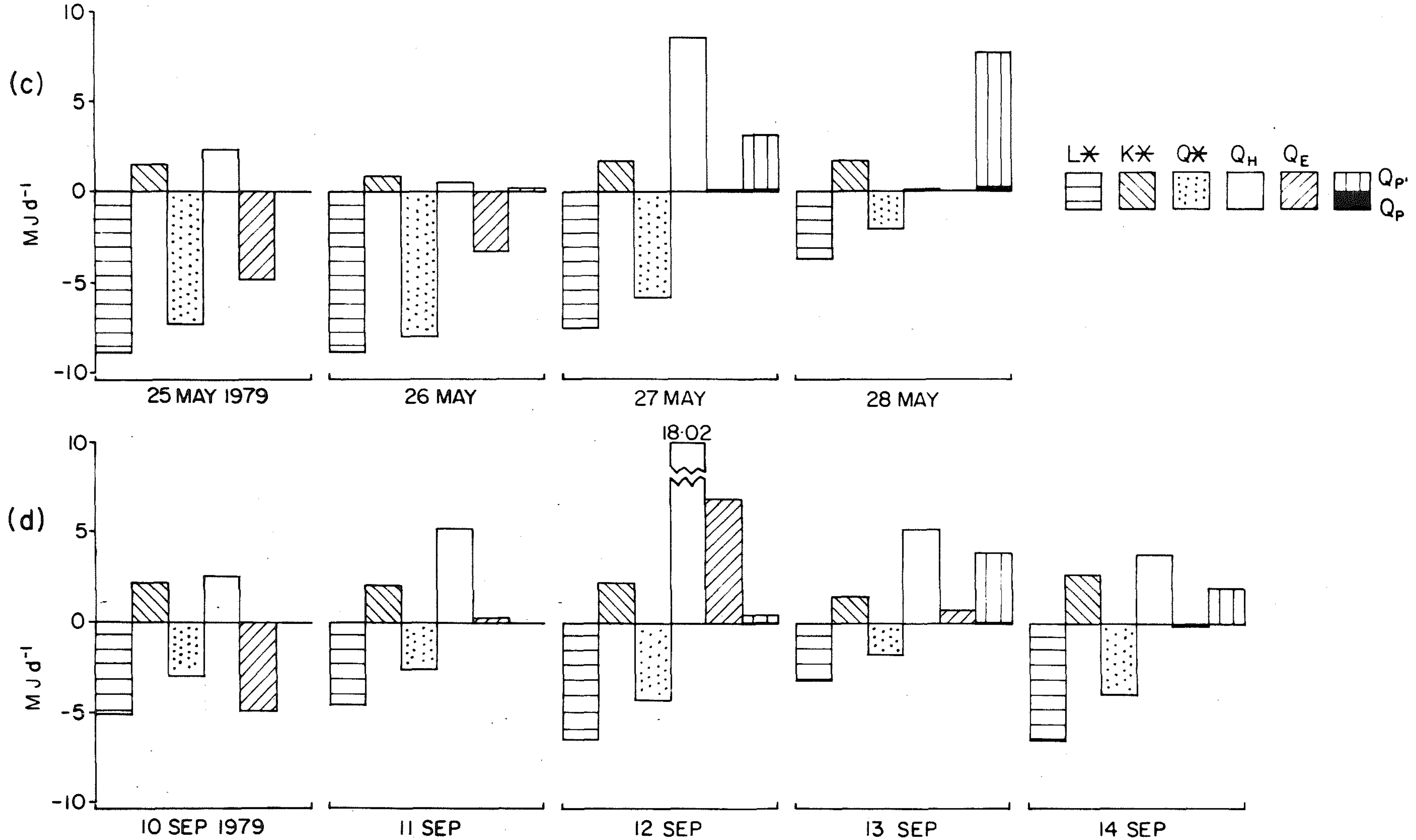
This report presents both components of Q_E (condensation-evaporation) and Q* (K* - L*) and in the calculation of percentage heat inputs condensation is treated as a separate heat supply. Heat flows are treated as positive towards the snowpack and negative away. The total, average and maximum snow depth decreases for each period are also listed in table 6.3.

The most striking feature of the energy balance results is that net radiation is negative for all winter melt days. In all cases, the low levels of daytime solar radiation are counteracted by the daily total of effective radiation. However, this does not mean radiation is unimportant in melt production. On many days a positive radiation balance develops during the daylight hours and on a daily average does net radiation become negative. In comparing the relative importance of heat supplies to the snowpack, K* was used instead of Q*. The significance of this is discussed in the comparisons.

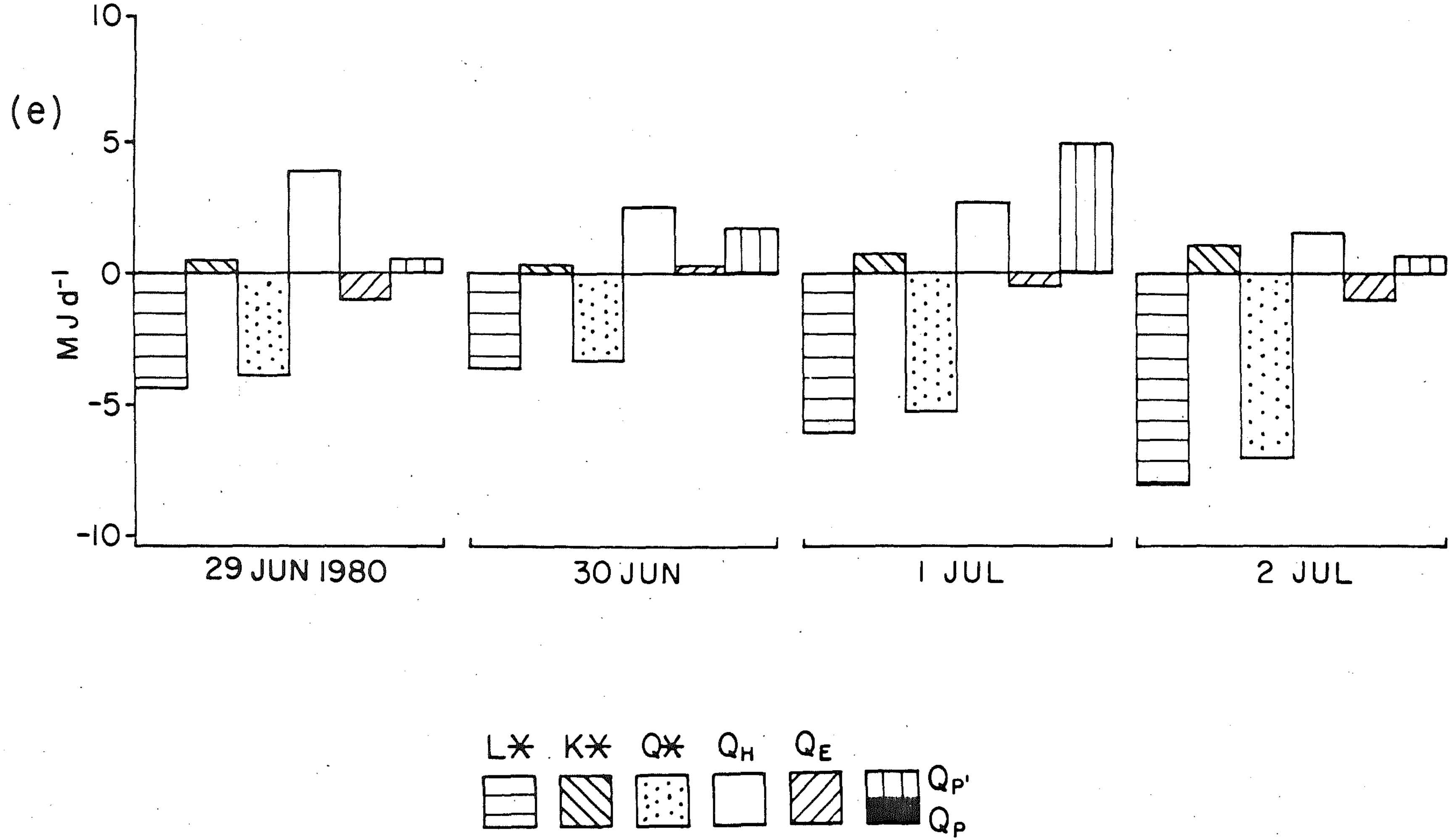
It is apparent from table 6.3 that the major heat flow for all winter melt periods was sensible heat, accounting for 65 to 78% of the total heat input. These values would rise to 81 - 96% if, as in normal practice, net radiation was used in the percentage calculations in

Figure 6.5 a-e DAILY HEAT FLOWS FOR THE WINTER SNOWMELT PERIODS.
(a) May 12 - 15, 1977; (b) May 21 - 23, 1977;
(c) May 25 - 28, 1979; (d) September 10 - 14, 1979;
(e) June 29 - July 2, 1980.


(a)
10
5
0
-5
-10
M J d⁻¹
12 MAY 1977
13 MAY
14 MAY
15 MAY
(b)
21 MAY 1977
22 MAY
23 MAY
L*
K*
Q*
Q_H
Q_E
$Q_{P'}$
Q_P



(c)
10
5
0
-5
-10
MJd-1
25 MAY 1979
26 MAY
27 MAY
28 MAY
L*
K*
Q*
QH
QE
QP'
QP
18·02
(d)
10
5
0
-5
-10
MJd-1
10 SEP 1979
11 SEP
12 SEP
13 SEP
14 SEP


(e)
10
5
0
-5
-10
MJ d⁻¹
29 JUN 1980
30 JUN
1 JUL
2 JUL
L*
K*
Q*
Q_H
Q_E
$Q_{P'}$
Q_P

MELT PERIOD	K^*	Q_H	Q_E	Q_P	SNOW DEPTH DECREASE (mm) TOTAL	AVERAGE	MAX
MAY 12-15, 1977	20.4	75.4	4.0	0.2	170	43	100
MAY 21-23, 1977	16.3	72.3	10.5	0.9	110	37	80
MAY 25-28, 1979	32.5	64.9	0.6	2.1	150	38	80
JUNE 29-JULY 2,1980	19.1	77.5	2.5	0.8	110	28	60
SEPTEMBER 10-14,1979	19.6	65.2	14.9	0.3	150	25	40

Table 6.3 PERCENTAGE HEAT INPUT AND SNOW DEPTH DECREASE - WINTER SNOWMELT PERIODS. K^* is used in place of Q^* because of the negative net radiation balance but Q_H still dominates. Average and maximum snow depth decrease are per day.

place of net shortwave radiation. Inputs of solar radiation were most important during the May 25-28, 1979 period which, as discussed, had air temperatures slightly below the monthly average and reasonably clear sky conditions. However, K^* during this period still only contributed 33% of the total heat input with Q_H contributing by far the majority at 65%. These results concur with the view propounded by the U.S. Army (1960) that during the winter, convection is the dominant heat supply.

Condensation, Q_E, also supplied appreciable quantities of heat to the snowpack averaging approximately 6% over the five periods. During the September period, it accounted for a total of 7.89 MJ or 15% of the total heat supply. This was also the only period which recorded an average vapour pressure in excess of that for a melting snow at 6.11 mb (table 6.2).

Precipitation heat flow produced only small amounts of energy, on only one occasion (table 6.3) exceeding 2% of the total energy input. The single greatest daily input was on July 1, 1980 when a 22 mm rainfall contributed 0.26 MJ to the snowpack. However, the possibility exists that precipitation may have contributed far greater quantities of heat if the snowpack had a heat deficit and the rain was frozen within the ice matrix. The release of latent heat, as outlined earlier, could considerably increase the relative importance of the precipitation heat flow. Values of Q_P', representing the additional heat supply from the release of latent heat, are included on figures 6.5a to 6.5c with the much lower values of Q_P underneath. On days of heavy rainfall, values of Q_P' exceed many of those for Q_E and, on occasion, even Q_H. In particular on May 28, 1979, if 22 mm of rainfall had been totally frozen within the snowpack the precipitation heat flow would rise from 0.3 to 7.6 MJ, 81% of the daily heat input. On this particular day inputs of sensible heat were small because of low windspeeds but even over the entire May 25 - 28, 1979 period Q_P' could have contributed 39% of the 17.85 MJ heat supply.

The degree to which rainfall was frozen within the snowpack is unknown, but the suspicion is that it was negligible The main reason is that the heat deficit of the snowpack has proven to be small, in many cases smaller than the heat which would be provided by a 10 mm rainfall being totally frozen within the snowpack. Since most rainfall was recorded at the end of warming trends, the heat deficit would likely have been negated or at least greatly reduced by the other heat sources prior to the rainfall. Hence, for most melt periods, the only precipitation heat flow is expected to come directly from the heat content of the rain.

6.6.4 Days of Greatest Heat Input

According to the snow depth decreases listed in table 6.4, there is one day in each of the four periods which recorded a snow depth decrease twice the average daily decrease. As noted earlier, although snow depth is not an accurate measure of melt, it is a good indicator of melting conditions. In view of this, it is interesting that the days which recorded maximum snow depth decreases, also had the largest heat inputs (table 6.4). In most cases the heat input on these days is appreciably greater than for other days in the same period. On two occasions, May 25-28, 1979 and September 10-14, 1979, the heat input from one day was over 50% of the total heat input recorded for the entire melt period.

As these particular days are so important to snowmelt, table 6.5 was constructed listing the relative importance of the four major heat inputs. Importantly, on all days some rainfall was recorded and cloudy conditions dominated. Although the precipitation heat flow percentages are still small, contribution from the latent heat of condensation

DATE	DAILY HEAT INPUT ($MJ\,d^{-1}$)	TOTAL HEAT INPUT (MJ)	MEAN HEAT INPUT ($MJ\,d^{-1}$)
MAY 12, 1977	6.36		
MAY 13, 1977	7.64		
MAY 14, 1977	+<u>11.03</u>		
MAY 15, 1977	5.31	30.34	7.59
MAY 21, 1977	5.28		
MAY 22, 1977	+<u>9.00</u>		
MAY 23, 1977	6.10	20.38	6.79
MAY 25, 1979	3.90		
MAY 26, 1979	1.40		
MAY 27, 1979	+<u>10.49</u>		
MAY 28, 1979	2.06	17.85	4.46
JUNE 29, 1980	+<u>4.42</u>		
JUNE 30, 1980	3.19		
JULY 1, 1980	3.51		
JULY 2, 1980	2.61	13.73	3.43
SEPTEMBER 10, 1979	4.66		
SEPTEMBER 11, 1979	7.50		
SEPTEMBER 12, 1979	+<u>27.04</u>		
SEPTEMBER 13, 1979	7.47		
SEPTEMBER 14, 1979	6.39	53.06	10.61

Table 6.4 TOTAL, MEAN AND DAILY HEAT INPUTS - WINTER MELT PERIODS.
'+' denotes days of greatest snow depth decrease; underlining refers to days of highest heat transfer.

	% OF ENERGY INPUT			
	K*	Q_H	Q_E	Q_P
MAY 14, 1977	12.8	75.9	11.1	0.3
MAY 22, 1977	9.3	65.1	23.9	1.7
MAY 27, 1979	16.1	81.9	1.0	1.0
JUNE 29, 1980	11.1	88.2	0.0	0.2
SEPTEMBER 12, 1979	8.1	66.7	25.2	0.1

Table 6.5 PERCENTAGE HEAT INPUT ON DAYS OF GREATEST HEAT SUPPLY.
These are the same days on which maximum snow depth decreases were observed. Some rainfall also occurred on each day.

averaged over 12% and for two periods exceeded 20%. However, the dominance of sensible heat transfer is still very pronounced, the average contribution of total heat input for the five periods being in excess of 75%.
In accordance with rainy, cloudy conditions, the percentage for net shortwave radiation ranged from only 8 to 16%. It must be remembered that this comparison employed K* in place of Q*. If Q* had been used the average percentage inputs from the sensible, latent and precipitation heat flows over the five days would be 85.7, 13.5 and 0.7 respectively.

6.6.5 Summary - Comments

Based on the above results, winter melt periods are apparently dominated by large inputs of sensible heat and a negative radiation balance. Inputs of shortwave radiation play only a minor role, although the largest melt rates are expected to be concentrated in the daylight hours when inputs from both sensible and radiation heat flow are maximized.

The greatest decreases in snow depth were recorded on days of rainfall and although the precipitation heat flow remained small, appreciable quantities of heat were received from condensation. However, sensible heat transfer still dominated.

There were two potential problems associated with the calculation of heat flows during the winter season. The first was associated with the ratios used to estimate incoming solar radiation at the SB site based on data collected at the CF site. The ratios for the winter months were the lowest for the entire year and showed the greatest scatter. As a result the estimates for the SB radiation receipts for the winter periods may be slightly inaccurate. However, even if K↓ was underestimated the size of the error is unlikely to be sufficient to produce a net positive radiation balance. Hence, sensible heat flow would remain as the dominant heat flow.

The second problem relates to the assumption of a melting snow surface for all calculations. During the nights, refreezing of the snow surface may have taken place. If the surface snow temperature decreased, values of L↑ and hence Q_R would be overestimated. However, surface cooling was reasoned to be small because most melt periods were characterized by high inputs of sensible heat and extensive low cloud cover.

A sub-freezing surface snow temperature would also have resulted in greater temperature and vapour pressure gradients between the air and snow. Values of Q_H and Q_E would have been increased accordingly. Hence, the net effect of underestimating the snow surface temperature would be to underestimate the transfers of sensible and latent heat and overestimate net radiation. In view of the overpowering importance of Q_H presented above, this potential error is not considered significant.

6.7 SPRING MELT PERIODS

6.7.1 General

For the study years 1975 - 1980, the periods of intense spring melt at the SB site all began in September or early October, except for 1976. In that year, intense and continuous melt did not begin until November and snowcover remained until early December (figure 6.4). In contrast, the years 1975, 1978, 1979 and 1980 all recorded the first total disappearance of snow in early November, a full month earlier than in 1976. No end of season record was available for 1977. Morris and O'Loughlin (1965) and O'Loughlin (1969a) report, for the same area over the years 1962-68, the disappearance of snow from as early as September to as late as January. The 1975-80 years would therefore appear to have more consistent dates for snowcover disappearance than might normally be expected.

The duration of snowcover depletion for the 1975-80 years was found to be approximately one month, although the steepness of the depletion curves in figure 6.4 noticeably increases for 1976 when final melt was delayed until late spring. This rapid depletion may be accounted for by the greater snow accumulation in 1976 and because melt rates are expected to increase in late spring with increases in available solar energy.

In an attempt to assess the approximate loss of water from the snowpack over the spring snowmelt periods, table 6.6 was constructed. Dates for the end of the major accumulation period and beginning of spring snowmelt were subjectively based on the slope of the accumulation line and the occurrence of late season snowfalls, as depicted in figure 6.4 and table 6.6.

YEAR	PERIOD OF MAIN SPRING SNOWMELT	SNOW DEPTH DECREASE (mm)	AVERAGE SNOW DEPTH DECREASE (mm d^{-1})	TOTAL # OF DAYS	TOTAL SPRING MELT (mm)
1975	SEPTEMBER 14-NOVEMBER 7	1730	32	54	692
1976	NOVEMBER 12-DECEMBER 8	1560	58	27	624
1977	- NO FINAL RECORD -				
1978	OCTOBER 24-NOVEMBER 20	1290	48	27	516
1979	OCTOBER 15-NOVEMBER 6	1200	57	21	480
1980	SEPTEMBER 23-NOVEMBER 5	1860	43	43	744

Table 6.6 DURATION AND SNOW DEPTH DECREASE DURING SPRING. Estimates of the total spring melt are based on the listed snow depth decrease and a mean snow density of 400 kg m^{-3}.

In general, snow depth is seen to decrease at a rate of approximately 32 to 58 mm d^{-1} averaging 48 mm d^{-1}. Assuming a mean melting snow density of 400 kg m^{-3} (which would account for lower densities at the beginning of melt and higher ones at the end, from figure 4.19) spring snowmelt would contribute 13 - 23 mm of runoff per day to the stream and/or underground water supplies. These rates are considerably higher than the range of 7 - 11 mm d^{-1} suggested by O'Loughlin (1969b) for the same area. However, they are not unreasonable in comparison to the 3 mm hr^{-1} and up to 9 mm hr^{-1} rates reported by Kuzmin (1961) for the European USSR. Over the entire spring melt period these melt rates contribute significant quantities of runoff, ranging from 0.54 to 0.83 m.

6.7.2 Description of Spring Snowmelt Periods

In an effort to quantify the relative importance and magnitude of the various heat inputs to spring snowmelt, specific periods were selected from the 1975-80 meteorological record, according to the methods outlined in section 6.4. Four periods were finally selected, one from each of the years 1976, 1977, 1979 and 1980. The latter three occurred in the month of October while the fourth period in 1976 was in November. The length of these periods ranged from five to fourteen days, the prevailing weather conditions for which are summarized in table 6.7.

For all the October periods, windspeeds and air temperatures were above the longterm mean monthly average.

	AIR TEMPERATURES (^{o}C)		MAXIMUM AIR TEMPERATURES (^{o}C)		WINDSPEED (m s^{-1})		GLOBAL RADIATION (MJ d^{-1})		VAPOUR PRESSURE (mb)
	MEAN	DIFF.	MEAN	DIFF.	MEAN	DIFF.	MEAN	DIFF.	MEAN
OCTOBER 26-30, 1977	7.3	+4.8	10.8	+4.9	4.5	+1.5	20.40	+3.17	5.76
OCTOBER 24-29, 1979	6.0	+3.5	8.9	+3.0	4.2	+1.2	17.59	+0.36	6.78
OCTOBER 22-29, 1980	7.4	+4.9	10.8	+4.9	3.4	+0.4	16.48	-0.75	6.61
NOVEMBER 15-28, 1976	3.8	-1.1	7.4	-1.2	2.7	-0.1	16.50	-3.54	6.36

Table 6.7 WEATHER CONDITIONS DURING SPRING SNOWMELT PERIODS. MEAN values refer to those calculated over the respective study period. DIFF. refers to the difference between the MEAN value and the long term mean monthly value.

Radiation receipts in 1979 and 1980 were reasonably consistent with the average, but the five day period in October 1977 recorded global radiation over 3 MJ d^{-1} or 18% in excess of the mean monthly value.

In contrast to the above, the November 15 - 28, 1976 period was characterized by temperatures and windspeeds below normal and reasonably cloudy periods such that receipts of shortwave radiation were 3.5 MJ d^{-1} below average. However, conditions were known to fluctuate during these lengthy periods. In general, all melt periods contained either one or more series of clear skies followed by increasing cloud and temperature. In most cases, the warmest air temperatures were associated with periods of increasing cloud and wind prior to the days of precipitation. Specific details of the weather conditions pertinent to snowmelt are discussed in the next section.

MELT PERIOD	TOTAL HEAT INPUT (MJ)	POTENTIAL MELT (mm of water)	CALCULATED DECREASE IN SNOW DEPTH (mm)	OBSERVED DECREASE IN SNOW DEPTH (mm)
OCTOBER 26-30, 1977	75.4	225	500	300
OCTOBER 24-29, 1979	76.4	228	507	500
OCTOBER 22-29, 1980	79.4	237	527	490
NOVEMBER 22-29, 1976	114.8	343	762	810

Table 6.8 CALCULATED AND OBSERVED SNOW DEPTH DECREASES. The calculated decreased are calculated based on the total heat input, latent heat of fusion and a snow density of 450 kg m^{-3}.

6.7.3 Accuracy of the Heat Flow Calculations

For the spring melt periods, it was possible to compare actual decreases in snow depth with those which would be expected from the total heat inputs. This type of comparison was not attempted for the winter period because of the expected large fluctuations in snow density. However, for the spring periods the assumption was made that the snow had reached a consistent melting snow density. This was set at 450 kg m^{-3} according to the results displayed in figure 4.19 and the discussion on melting snow densities in section 4.6.

Table 6.8 compares the actual decrease in snow depth recorded at the SB site with the expected decrease from inputs of radiation (Q*), sensible, latent and precipitation heat. For three of the spring melt periods, differences of less than 10% existed between the observed and calculated values. The differences noted for the October 26-30, 1977 period can be partially explained by large snow depth decreases which occurred between the times of snow depth measurement on October 25 and 26. An evaluation of the heat flows for the 25th could not be made because of missing data. The amount of melt which may have occurred in the morning of the 26th prior to the mid-day snow depth measurement is unknown. However, there is a strong possibility that an appreciable decrease in snow depth did occur and which is not accounted for in table 6.8.

In general, the heat flow calculations are considered to produce results in good agreement with those for the observed snow depth decreases.

6.7.4 Heat Flow Results

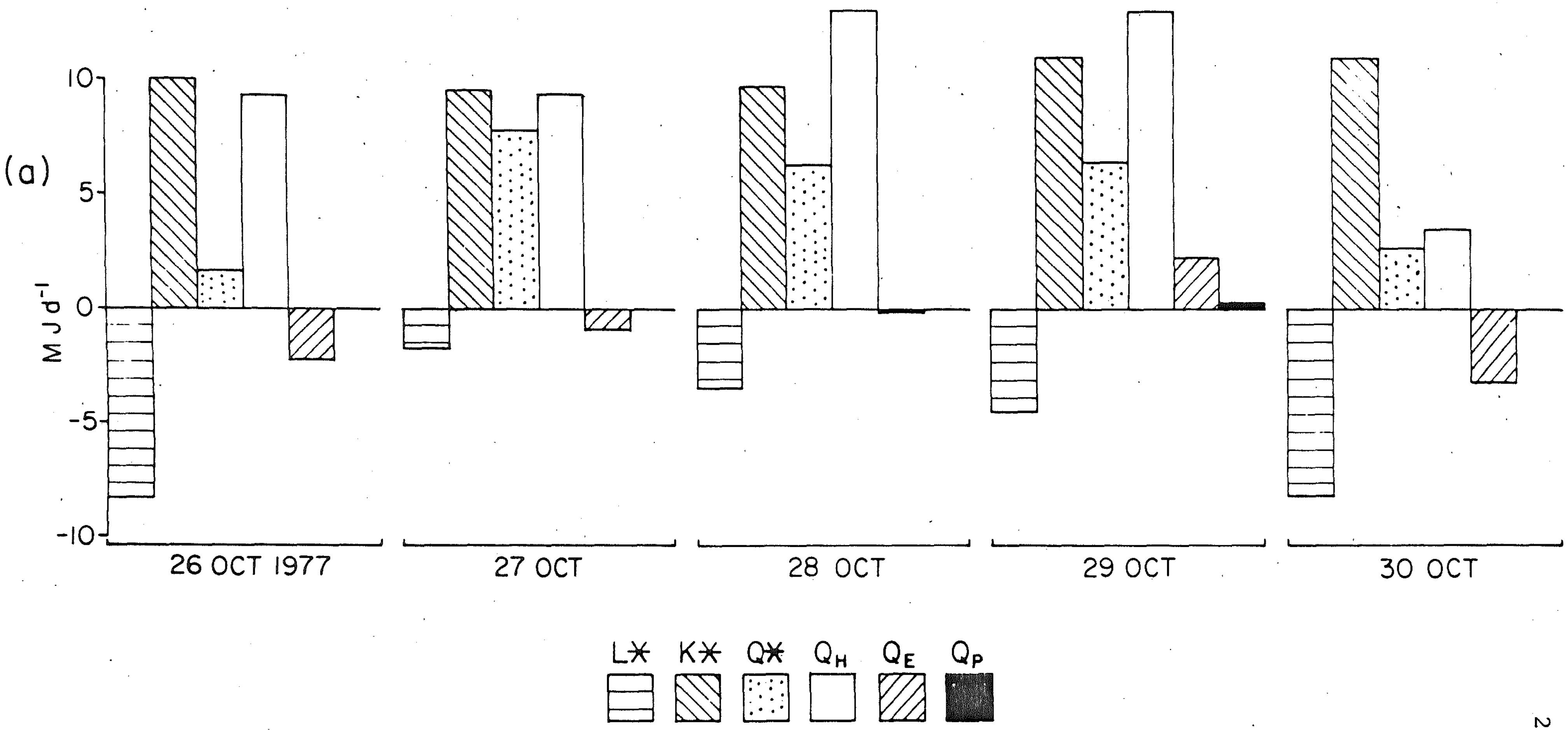
The calculated heat flows for each of the melt periods are illustrated in figures 6.6a to 6.6d. Summaries of the heat inputs based on both K* and Q* as the single radiation input are presented in table 6.9.

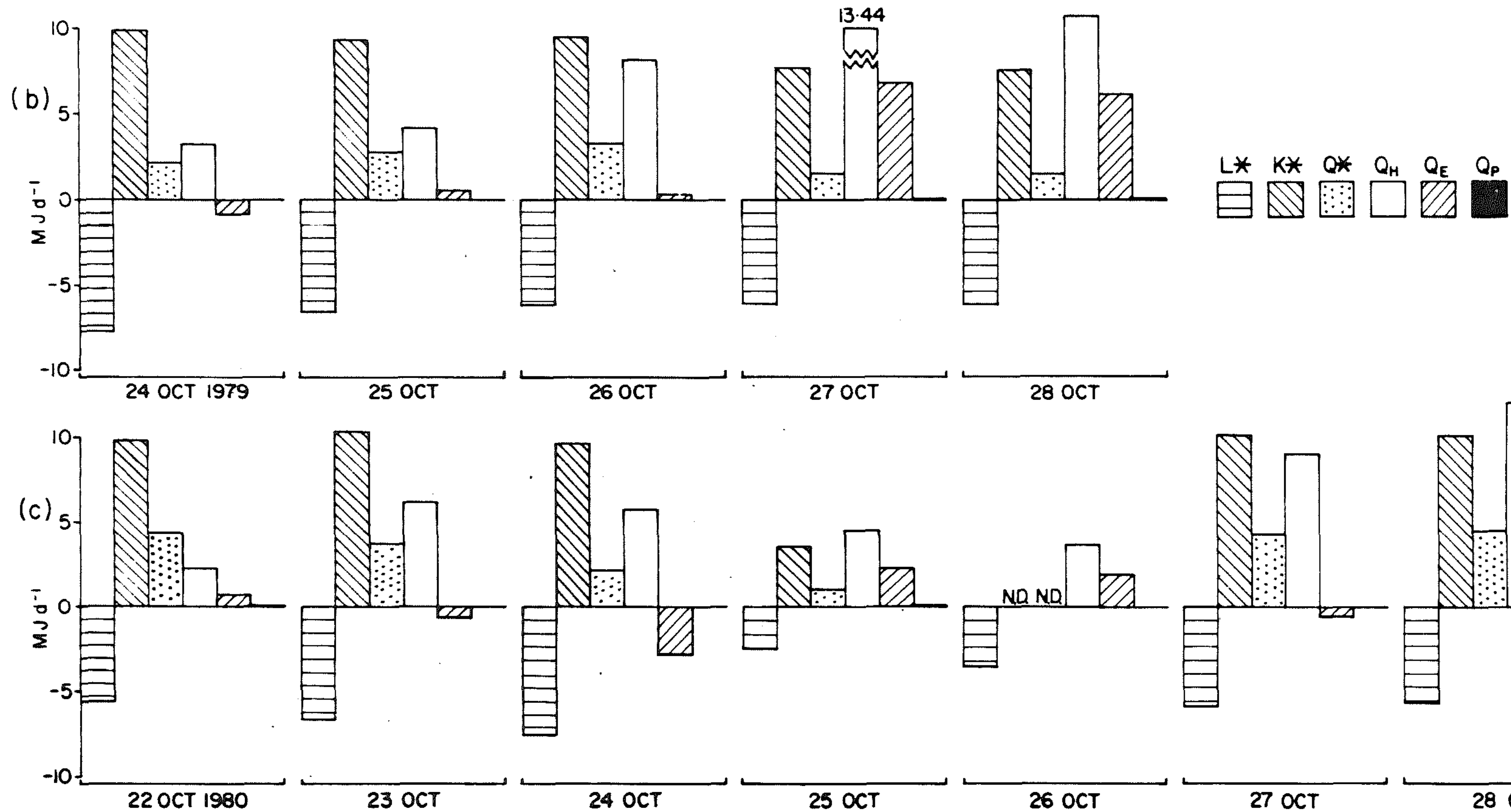
As expected, these periods were characterized by a positive radiation balance. While this can mainly be attributed to the seasonal increase in solar radiation, on some occasions, it resulted from significant decreases of L* in warm cloudy weather. Similar results have been found by other researchers such as Hoinkes (1970) in the Antarctic; Holmgren (1971) in the Arctic, who found during summer that dense cloud doubled the net radiation balance of clear days; and, Sauberer and Dimhirm (1952) in the European Alps, who reported a negative Q* value during clear skies but a high positive value during complete cloud cover.

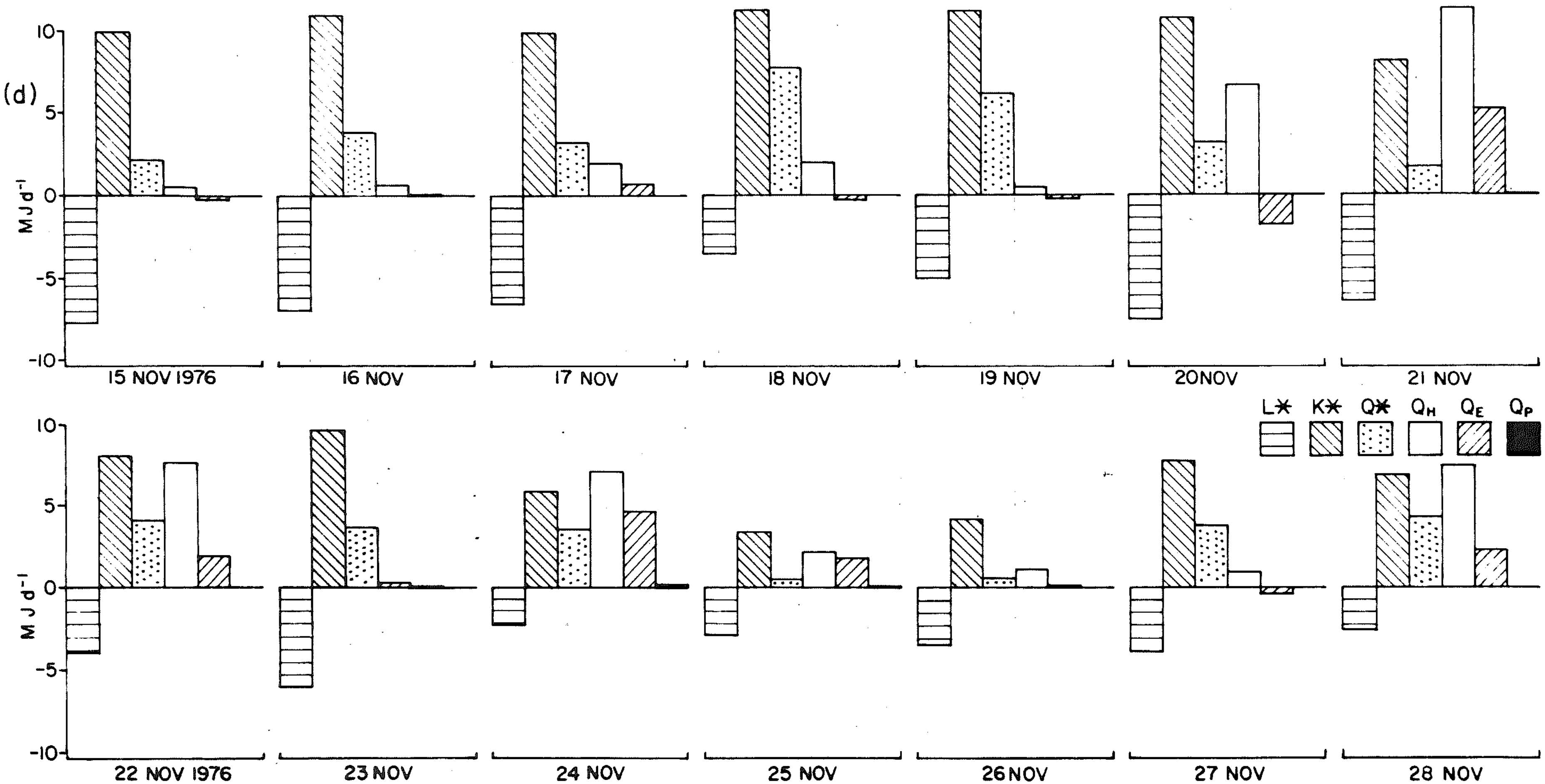
Table 6.9a considers the percentage inputs of heat from the four major heat flows, and as for section 6.6, treats K* and L* as separate flows. Based on these calculations, net solar radiation can be seen to be the major heat component for all spring melt periods, accounting for 45 to 64% of the total heat input. K* is especially important during the November period which can partially be attributed to increases in available solar radiation between October and November, but more importantly because of low inputs from sensible heat transfer. As mentioned, the November period was dominated by relatively low windspeeds and below normal temperatures.

A more common method of presenting energy balance

Figure 6.6 a-d DAILY HEAT FLOWS FOR THE SPRING SNOWMELT PERIODS.
(a) October 26 - 30, 1977; (b) October 24 - 28, 1979;
(c) October 22 - 29, 1980; (d) November 15 - 28, 1980.


(a)
MJ d⁻¹
10
5
0
-5
-10
26 OCT 1977
27 OCT
28 OCT
29 OCT
30 OCT
L*
K*
Q*
Q_H
Q_E
Q_P


13·44
(b)
MJ d⁻¹
10
5
0
-5
-10
24 OCT 1979
25 OCT
26 OCT
27 OCT
28 OCT
L* K* Q* Q_H Q_E Q_P
(c)
22 OCT 1980
23 OCT
24 OCT
25 OCT
26 OCT
ND ND
27 OCT
28 OCT
29 OCT


(d)
10
5
0
-5
-10
MJ d⁻¹
15 NOV 1976
16 NOV
17 NOV
18 NOV
19 NOV
20NOV
21 NOV
L* K* Q* QH QE QP
22 NOV 1976
23 NOV
24 NOV
25 NOV
26 NOV
27 NOV
28 NOV

	PERCENTAGES OF HEAT INPUT BASED ON TOTALS USING:										
	a) K*				b) Q*						
									SNOW DEPTH DECREASE		
MELT PERIOD	K*	Q_H	Q_E	Q_P	Q*	Q_H	Q_E	Q_P	TOTAL	DAILY MEAN	DAILY MAX
NOVEMBER 15-28, 1976	63.8	27.0	9.0	0.2	42.0	43.4	14.4	0.3	810	58	120
OCTOBER 26-30, 1977	50.2	47.3	2.2	0.4	32.8	63.7	3.0	0.5	300	60	160
OCTOBER 24-29, 1979	45.0	40.7	14.2	<0.1	17.2	61.3	21.3	0.1	500	83	150
OCTOBER 22-29, 1980	51.6	39.6	8.2	0.5	26.9	59.8	12.4	0.8	490	61	100
AVERAGE	52.7	38.7	8.4	0.3	29.7	57.1	12.8	0.4			

Table 6.9 a,b PERCENTAGE HEAT INPUT AND SNOW DEPTH DECREASE - SPRING SNOWMELT PERIODS. The first four columns of results (a) are based on net solar radiation, K*, in the percentage calculations, while in (b) net radiation, Q*, was employed.

results employs Q^* rather than K^* in the calculation of percentage heat inputs. If this convention is used (table 6.9b), Q_H provides the greatest heat input in each melt period. The overall average for the four periods is 57% but, for the reasons explained above, in November the inputs from Q_H and Q^* almost balance.

The contribution of heat from condensation varied considerably from 3.0 to 21.3%, averaging 12.8% over the four periods. In view of the various results reviewed by Paterson (1969), only the October 1979 value of 21.3% can be considered as significantly high.

Precipitation heat flow contributed very small amounts of energy, consistently less than 1% for all periods. No estimates were made for Q_P' as rainfall was extremely unlikely to be refrozen within the snowpack during spring melt conditions.

Although the average percentage contribution of heat from precipitation was small, the highest daily heat input for each period occurred on days of rainfall. With the exception of October 1977, these days were found to be the same days which recorded the maximum decreases in snow depth (table 6.10). In the case of October 1977, the maximum decrease was recorded on October 28, one day before the day of rainfall, but both days had comparable total heat flows (19.20 MJ and 21.89 MJ).

6.7.5 Days of Greatest Heat Input

On average, 30% (range 15 - 53%) of the total heat

DATE	DAILY HEAT INPUT (MJ d^{-1})	TOTAL HEAT INPUT (MJ)	MEAN HEAT INPUT (MJ d^{-1})
OCTOBER 26, 1977	10.96		
OCTOBER 27, 1977	17.09		
OCTOBER 28, 1977	+ 19.20		
OCTOBER 29, 1977	R 21.89		
OCTOBER 30, 1977	6.21	75.35	15.07
OCTOBER 24, 1979	5.43		
OCTOBER 25, 1979	7.51		
OCTOBER 26, 1979	11.71		
OCTOBER 27, 1979	R + 21.82	64.93	12.99
OCTOBER 28, 1979	18.46		
OCTOBER 22, 1980	R 7.18		
OCTOBER 23, 1980	9.83		
OCTOBER 24, 1980	7.79		
OCTOBER 25, 1980	R 7.66		
OCTOBER 26, 1980	-		
OCTOBER 27, 1980	13.25		
OCTOBER 28, 1980	R + 21.35		
OCTOBER 29, 1980	R 6.85	73.92	9.24
NOVEMBER 15, 1976	2.73		
NOVEMBER 16, 1976	4.56		
NOVEMBER 17, 1976	5.72		
NOVEMBER 18, 1976	9.65		
NOVEMBER 19, 1976	6.54		
NOVEMBER 20, 1976	9.79		
NOVEMBER 21, 1976	R + 18.07		
NOVEMBER 22, 1976	13.65		
NOVEMBER 23, 1976	4.03		
NOVEMBER 24, 1976	R 15.37		
NOVEMBER 25, 1976	R 4.43		
NOVEMBER 26, 1976	R 1.73		
NOVEMBER 27, 1976	4.71		
NOVEMBER 28, 1976	13.84	114.82	8.20

Table 6.10 TOTAL, MEAN AND DAILY HEAT INPUTS - SPRING SNOWMELT PERIODS
'+' denotes days of greatest snow depth decrease; underlining refers to days of highest heat transfer; 'R' indicates days on which some rainfall was recorded. Note that the days of highest heat transfer all coincide with days of rainfall.

input for the four spring melt periods occurred on the high heat input days noted in table 6.10. The relative importance of the various heat flows for these specific days appear in table 6.11.

Comparing the average percentage results from table 6.9b and 6.11 the importance of net radiation noticeably decreases by approximately 50% on days of high heat input. The latent heat flow from condensation almost totally accounts for the difference rising to 23.2% or almost twice the mean overall value. Inputs from sensible heat remained relatively constant near the 60% mark and although all of these days recorded some precipitation, Q_P increased by only 0.1%.

	PERCENTAGE OF TOTAL HEAT INPUT			
	Q^*	Q_H	Q_E	Q_P
OCTOBER 29, 1977	29.1	59.0	10.2	1.6
OCTOBER 27, 1979	7.0	61.6	31.3	0.2
OCTOBER 28, 1980	20.8	56.3	22.8	0.1
NOVEMBER 21, 1976	9.2	62.1	28.6	0.2
AVERAGE	16.5	59.8	23.2	0.5

Table 6.11 PERCENTAGE HEAT INPUTS ON DAYS OF GREATEST HEAT SUPPLY - SPRING SNOWMELT PERIODS. As for the entire melt periods Q_H dominates but Q_E is in three cases greater than Q^*.

6.7.6 Days of Maximum Rainfall

The relatively small values of Q_P described above are in direct contrast to the results of Fitzharris et al. (1981), who report that precipitation heat flow may contribute up to 20% of the total heat input during a rain on snow event. In order to pursue this aspect further, all days which recorded large inputs (> 5 mm d^{-1}) of rainfall were analyzed for the relative importance of the main heat flows (table 6.12). In general, the relative importance of Q*, Q_H and Q_E remained approximately the same as for the days of high heat input. Minor decreases in Q_H and Q_E were taken up by the precipitation heat flow but it averaged only 3%, with a maximum of 7.8% on October 29, 1980.

The main reason for the difference between these results for precipitation heat flow and those offered by Fitzharris et al.(1981) and Anderton and Chinn (1978) (described in section 6.2) relates to the magnitude and temperature of the rainfall events. Fitzharris et al. 1981) were studying an exceptionally intense rainfall characterized by very warm temperatures. They refer to a 25 hour period of rain with intensities of 10 mm h^{-1} and rain temperatures ten degrees above freezing. In contrast, the size and temperature of rainfall events presented in this study are much smaller and lower. The maximum daily rainfall was only 22 mm and the approximate rain temperature only 5.5^{o}C above freezing. For the Craigieburn Mountains these conditions are probably more typical of rain on snow events and those reported by Fitzharris et al. (1981) should be considered representative of a rare event.

	PERCENTAGE OF TOTAL HEAT INPUT				TEMPERATURE AND AMOUNT OF RAINFALL	
	Q^*	Q_H	Q_E	Q_P	(mm d^{-1})	(^oC)
OCTOBER 29, 1977	29.1	59.0	10.2	1.6	19.0	5.5
OCTOBER 29, 1980	3.6	68.1	20.4	7.8	22.0	4.4
NOVEMBER 24, 1976	23.2	45.8	29.9	1.1	11.5	3.4
NOVEMBER 25, 1976	10.4	47.9	40.2	1.6	6.8	2.4
AVERAGE	16.6	55.2	25.2	3.0	14.8	3.9

Table 6.12 PERCENTAGE OF HEAT INPUT AND PRECIPITATION CHARACTERISTICS ON DAYS OF MAXIMUM RAINFALL. The temperature of rainfall was derived from measures of the wet bulb air temperature.

6.7.7 Summary

As evinced for the winter melt periods, sensible heat flow dominated (57%) the total heat input to the snowpack during spring snowmelt. Solar radiation was frequently the largest heat component but net radiation was secondary to Q_H, on average contributing only 30% of the total heat input. This result is in direct contrast to the frequently held view (U.S. Army 1960) that for spring snowmelt Q^* is always the dominant heat source and Q_H is of only secondary importance.

The days of greatest heat input were all found to have recorded some rainfall but precipitation heat flow was found never to exceed 8% of the total. The major differences in the relative importance of the heatflow on days of rainfall occurred in Q^* and Q_E with the percentage input from Q_H being similar to that for over the entire melt period. On days of heavy rainfall and high heat input, values for condensation heat flow actually exceeded those for net radiation.

Although the spring melt period is normally spread over a month or more, appreciable quantities of meltwater may be produced during intense melt periods. Heat contributed on some of the high input days is sufficient to melt over 65 mm d^{-1} of water from the snowpack.

6.8 DAYS OF HIGH SENSIBLE HEAT TRANSFER

Because transfers of sensible heat were found to be the most important heat transfer during snowmelt in both winter and spring, a decision was made to analyze the seasonal trend in high daily values of Q_H. Strong winds and warm air temperatures are both required to produce high levels of Q_H. Therefore, days were selected from the May to November periods 1975-1979, if they recorded:

a) a mean temperature greater than the longterm mean monthly maximum,

and, b) an average daily windspeed equal to or greater than 5 m s^{-1}. In total 74 days were identified, the results for which, based on a monthly separation, appear in table 6.13 and figure 6.7. The reported frequency of days per

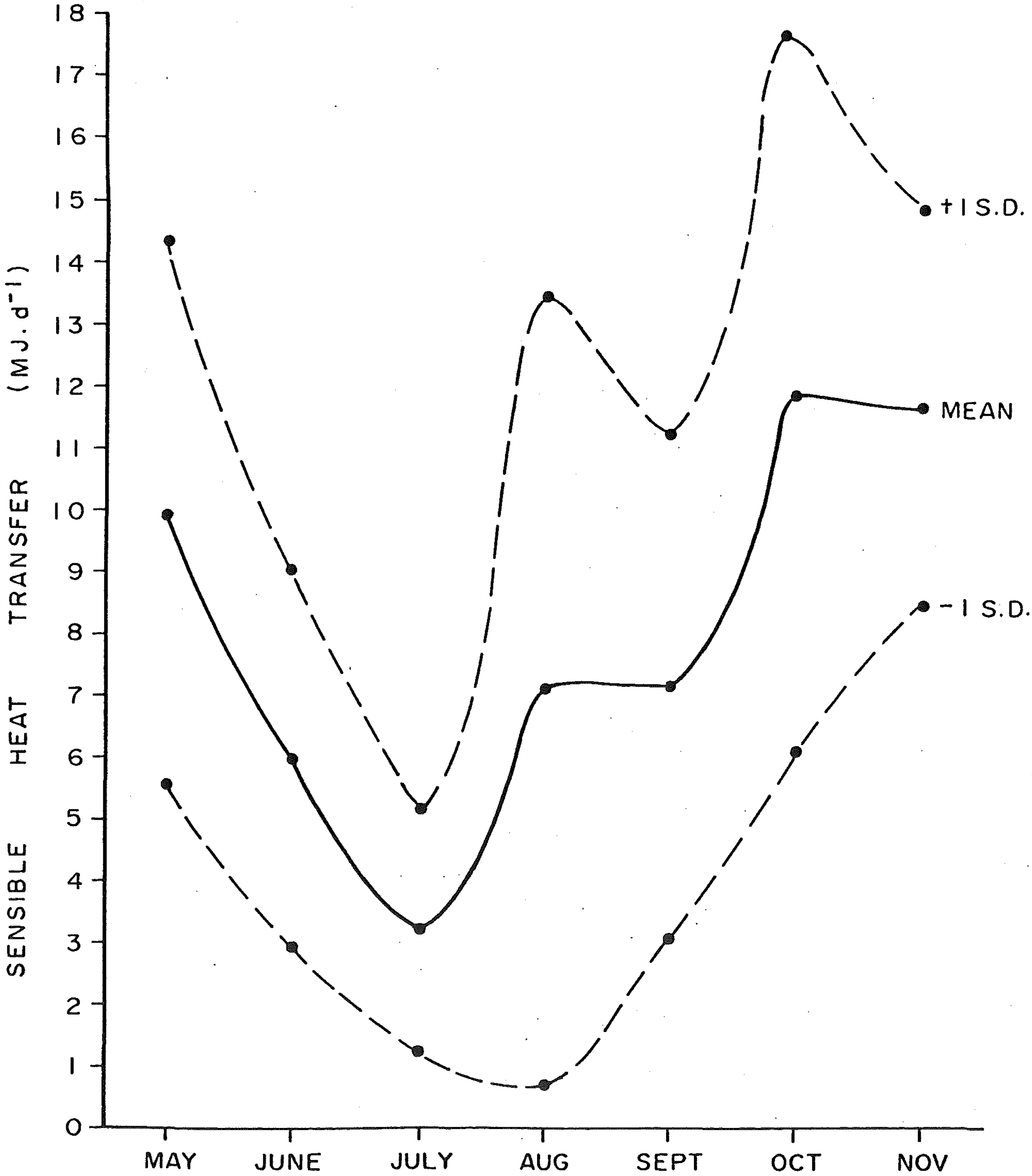


Figure 6.7 SEASONAL TREND IN MONTHLY AVERAGES OF SENSIBLE HEAT TRANSFER ON DAYS OF HIGH WINDSPEED AND AIR TEMPERATURE.
The dashed lines represent one standard deviation about the mean.

month cannot be considered an accurate indication of actual frequency because of the length of record, and missing data.

	SAMPLE SIZE	MEAN Q_H ($MJ\,d^{-1}$)	MAXIMUM Q_H ($MJ\,d^{-1}$)	MINIMUM Q_H ($MJ\,d^{-1}$)	MAXIMUM ERROR OF ESTIMATE 95% CONFIDENCE LEVEL
MAY	8	9.95	20.11	6.69	3.66
JUNE	7	5.98	9.59	2.57	2.82
JULY	11	3.20	6.54	0.51	1.33
AUGUST	11	7.07	21.95	0.46	4.28
SEPTEMBER	16	7.13	18.02	1.09	2.18
OCTOBER	15	11.84	22.63	5.40	3.19
NOVEMBER	6	11.63	16.75	7.80	3.14

Table 6.13 MONTHLY AVERAGES OF SENSIBLE HEAT TRANSFER ON DAYS OF HIGH WINDSPEED AND AIR TEMPERATURE.

As might be expected, there is a definite seasonal trend in the mean daily values which closely mirrors that for solar radiation (figure 2.4). The main discrepancy is that the lowest radiation values occur in June while those for Q_H are in July. Although these values are based on relatively small sample sizes, they are reasonably accurate estimates as evinced by the low maximum error of estimate figures.

The figures in table 6.13 compare quite favourably with results from other environments where sensible heat transfer is known to be important to snowmelt. Golding (1978) reports that during winter Chinooks in Canada, sensible heat contributed on average 3.4 MJ d^{-1} in 1975 and 5.67 MJ d^{-1} in 1976 to the snowpack. For the comparable timespan in table 6.13 (July, August and September), the main daily input is 5.80 M J. Interestingly, for 65% of the days used in the analysis, the main airflow was from the northwest (figure 6.8). This concurs with the freezing level data described in chapter three and supports the view that the warmest and strongest winds for the Craigieburn Mountains come from the northwest.

The maximum values for sensible heat transfer are also of interest. In general, they are approximately twice their respective mean monthly value and except for June and July are well in excess of 15 MJ d^{-1}. This analysis was based on the assumption that the surface snow temperature was at the melting point, while on some days, especially during the winter months, it may have been colder than $0^{o}C$. In such cases, the temperature gradient in equation 5.21 and the calculated values for Q_H would be underestimated. Apparently, sensible heat is capable of supplying large amounts of energy to the snowpack at any point in the snow season. However, the main question is whether these receipts would consistently exceed the inputs from net radiation. Based on the results in sections 6.6 and 6.7 Q_H appears to dominate Q^* at least for the selected periods of melt.

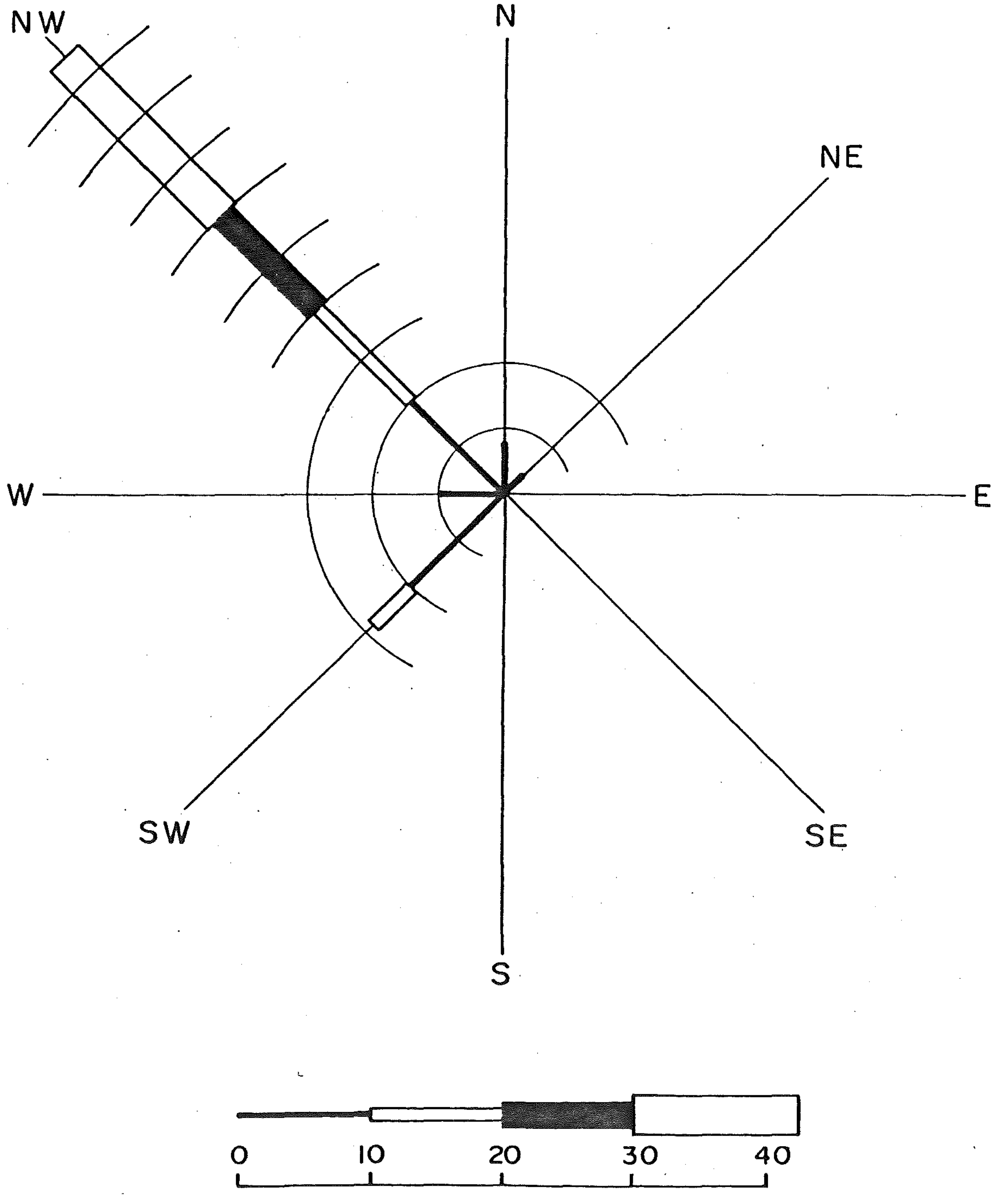


Figure 6.8 UPPER AIR WIND DIRECTION ON DAYS OF HIGH SENSIBLE HEAT TRANSFER.
Wind direction was obtained from the 800-700 mb level at Christchurch Airport.

6.9 EVAPORATION

6.9.1 General

Ablation of a snowcover does not necessarily have to occur through the process of melt and subsequent runoff. Evaporation from the snow surface to the overlying atmosphere may also take place, but the overall importance of evaporation compared to melt is a point of contention. Some workers, for example Matthes (1934), report entire snowfields ablating without contributing any meltwater to runoff, while others, as cited by Williams (1959), conclude that evaporation is seldom significant compared to melt. Over extended periods, the net moisture exchange frequently balances, or as observed by Martinelli (1960), will be a gain in the form of condensation.

Although evaporation has normally been found to be of secondary importance to melt, in the total ablation of a snowcover, there are specific weather conditions in which appreciable quantities of snow may be lost to evaporation. Most high evaporation rates have been reported during föhn or Chinook wind conditions [Ives (1950); Golding (1978); Bergen and Swanson (1964); Louie (1977)]. This section deals with the evaluation of evaporation rates for the Craigieburn Range under similar conditions.

6.9.2 Background to Evaporation

Evaporation from snow essentially involves a change from solid to vapour, the intervening liquid state usually only occurring if the surface of the snowpack is already at the melting point. Evaporation can take place only if there is a vapour pressure gradient between the snow surface and the overlying air. As described earlier, the vapour pressure over a melting snow surface can be said to be equal to the saturation vapour over ice at $0^{o}C$ which is 6.11 mb. If the vapour pressure of the air is less than this value, evaporation takes place, and if greater, condensation results. In order for evaporation to continue the air above the snowpack must be continually mixing, otherwise an equilibrium in vapour pressures would be reached and evaporation would cease.

When radiation is the major positive heat component of the energy balance, evaporation and condensation are known to follow a diurnal cycle [Meiman and Grant (1974); Martinelli (1960); Lemmela and Kuusisto (1974)]. Evaporation occurs during the day when the net radiation balance is maximized by inputs of solar radiation, and can provide the large quantities of energy required by the latent heat of vapourization. During the night a negative radiation balance frequently develops, resulting in decreases of the surface snow temperature and hence surface vapour pressure. If surface cooling is large enough, a reversal in the vapour pressure gradient will occur (Croft 1944) and the net vapour exchange will be positive towards the snow surface.

In most cases, the daily cycle of evaporation and condensation will balance or at best produce only small quantities of evaporation. However, as propounded by Diamond (1953) and Wilson (1941) the largest evaporation rates should occur during periods of low humidity and high sensible heat transfer. Under such conditions evaporation may take place, even at night, if the sensible heat transfer is sufficient to satisfy the latent heat of vapourization. The föhn type of wind, described previously, is characterized by high sensible heat transfer and low humidities and hence is ideal for snow evaporation. Ives (1950) reports a loss of 200 mm of water from a snowpack during one Chinook. Various other researchers as listed in table 6.14 have reported potential evaporation rates of over 2 mm d^{-1} during Chinooks. Storr (1968) estimated that a maximum rate of 13 mm d^{-1} is possible under föhn conditions.

6.9.3 Methods of Measurement

Three methods have regularly been used to determine evaporation from snow. These include:

a) volumetric

b) energy balance - Bowen Ratio

c) mass transfer

techniques.

a) The volumetric estimation of snow evaporation involves inserting pans or lysimeters into the snow surface and periodically weighing the containers to measure changes in the snow mass. Slaughter (1970) reviews the development, improvements and problems associated with pan measurements,

INVESTIGATOR	LOCATION	RESULTS
ANDERSON (1960)	CALIFORNIA U.S.A.	23 mm per season in forest 43 mm per season in clearing
BERGEN (1963)	COLORADO U.S.A.	1.0 - 2.0 mm d^{-1} over winter season average = 2.9 mm d^{-1}
BERGEN AND SWANSON (1964)	COLORADO U.S.A.	0.6 - 1.8 mm d^{-1}
CHURCH (1934)	LAKE TAHOE SIERRA NEVADA U.S.A. (1898 m)	26-44 mm mh^{-1} DECEMBER-APRIL average = 35 mm mh^{-1}
	MARLETTE LAKE SIERRA NEVADA U.S.A. (2438 m)	13-62 mm mh^{-1} DECEMBER-APRIL average = 34 mm mh^{-1}
	MT. ROSE SIERRA NEVADA U.S.A. (3292 m)	0.1 - 7.7 mm d^{-1} DECEMBER-JUNE average = 2.1 mm d^{-1}
	GREENLAND	NET AVERAGE = 35 mm mh^{-1} MAXIMUM = 85 mm mh^{-1} including föhn conditions
CROFT (1944)	WASATCH PLATEAU UTAH, U.S.A.	0.3 - 1.4 mm in 10 hour daylight period TRACE - 1.2 mm at night AVERAGE = 1.3 mm d^{-1}
GOLD AND WILLIAMS (1961)	OTTAWA CANADA	AVERAGE = 2.4 mm d^{-1} over 14 DAYS
GOLDING (1978); GOLDING AND SWANSON (1978)	ROCKY MOUNTAIN FOOTHILLS CANADA	1.2 - 2.0 mm d^{-1} during Chinooks
GUY (1936)	BOGONG HIGH PLAINS AUSTRALIA	4.3 - 9.1 mm mh^{-1} JUNE-OCTOBER AVERAGE = 6.5 mm mh^{-1}
HORTON (1934)	ROCKY MOUNTAINS COLORADO, USA	TOTAL WINTER ≃ 150 mm
HUTCHISON (1966)	ROCKY MOUNTAINS COLORADO, USA (2743 m)	0.4 - 1.2 mm d^{-1} APRIL 14-MAY 9 AVERAGE = 0.9 mm d^{-1}

IVES (1950)	COLORADO HIGH PLAINS, COLORADO U.S.A.	200 mm during one Chinook
LEMMELA AND KUUSISTO (1974)	SOUTHERN FINLAND (60 m)	0.39 mm d^{-1} during MARCH-APRIL 0.26 mm during DAYLIGHT 0.03 mm during NIGHT
MARTINELLI (1960)	FRONT RANGE COLORADO USA (3560 m)	0.68 mm d^{-1} during JULY
MEIMAN AND GRANT (1974)	PINGREE PARK ROCKY MOUNTAINS COLORADO, USA (2740 m)	135 mm TOTAL DECEMBER-APRIL 1.1 mm d^{-1} AVERAGE 3.2 mm d^{-1} MAXIMUM (for open forest sites)
MILLER (1955)	SIERRA NEVADA U.S.A.	2-3 mm d^{-1} MAXIMUM
WEST (1959; 1962)	SIERRA NEVADA U.S.A.	10-38 mm JANUARY-JUNE 50 mm per season MAXIMUM (forest and clearings)
WEST AND KNOERR (1959)	SIERRA NEVADA U.S.A.	2.3 - 12.7 mm mh^{-1} FEBRUARY-MAY 5.6 mm mh^{-1} AVERAGE (forest clearing)
WILLIAMS (1959)	OTTAWA CANADA	.001 - .068 mm hr^{-1} DECEMBER-FEBRUARY .026 mm d^{-1} AVERAGE

Table 6.14: RESULTS OF OTHER INVESTIGATIONS INTO SNOW EVAPORATION. A variety of methods were used by the various researchers. Short term rates have been included whenever possible.

and sets out a set of criteria for their use originally outlined by Sabo (1956).

Snow lysimeters were constructed for this study and are illustrated in figure 6.9. They were designed to limit problems of vapour pressure, meltwater collection and radiation attenuation but proved unsatisfactory in field tests. Strong winds in the alpine regions of the Craigieburn Mountains tended to transport as much snow in and out of the container as would be expected to occur from evaporation. A balance in wind transported snow could not be assumed nor could the pans be wind-shielded because any disturbance in the windflow would affect evaporation rates.

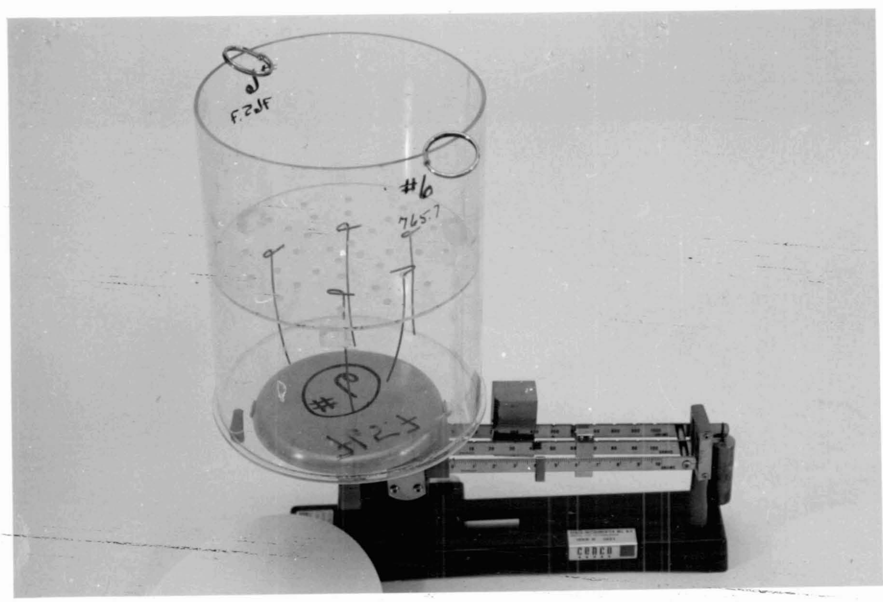


Figure 6.9 SNOW EVAPORATION PAN-LYSIMETER. Plexiglass construction was used to limit the effects of radiation absorption and tension devices were placed in the false bottom to allow free draining of the snow sample. Meltwater collected in the base of the pan and could be drained and weighed separately.

Since the greatest evaporation rates were expected during windy periods, this method was abandoned. The absence of documentation on the above problems in published works suggests most pan measurements have been conducted in relatively calm or sheltered environments.

b) The energy balance method is based on equation 5.1, whereby all but Q_E is measured or calculated and Q_E is treated as a residual. A second approach is to use the Bowen ratio (Bowen 1926) whereby both Q_H and Q_E are treated as residuals in the energy balance equation and the Bowen ratio is used to proportion the energy between the two. However, the validity of the use of the Bowen ratio for the accurate estimation of evaporation from snow is still very much a point of contention [Williams (1961); Storr (1979 pers comm)].

c) The third method of calculating snow evaporation is to evaluate the vapour flux between the atmosphere and snow surface. There are primarily two ways of calculating the flux which have been described in chapter five:

i) the eddy flux approach which relies on very sophisticated equipment.

ii) the aerodynamic approach, on which equations 5.21 and 5.22 are developed.

Based on derivations similar to that for Q_E, the equation for calculating the amounts of potential evaporation is [after Sverdrup (1936)]:

$$E = \frac{\gamma \rho_a k^2 u_z (e_a - e_s)}{P_a \ln (\frac{z'_a}{z_b}) \ln (\frac{z'_b}{z_o})} \qquad (6.1)$$

where E = potential evaporation (kg m^{-2} d^{-1}) and all other variables as defined for equations 5.21 and 5.22. Variations of this equation, primarily in reduced forms, are offered by Boyd (1967), Diamond (1953), Golding (1978), Light (1941) and Williams (1961).

6.9.4 Objectives and Methodology

The main objective of this section was to calculate daily rates of evaporation which according to Diamond (1953) and Light (1941) are at a maximum during periods of high sensible heat transfer.

The data set used for this analysis was the same as for section 6.8; all days where the mean temperature exceeded the longterm mean monthly maximum and the daily windspeed exceeded 5 m s^{-1}. Unfortunately, complete, accurate humidity records were available for only 46 of the 73 days. Temperatures and humidities were extracted from the chart records as previously outlined in this chapter.

6.9.5 Results

Although relative humidity values were generally high, averaging 74% over the 46 days (figure 6.10a),

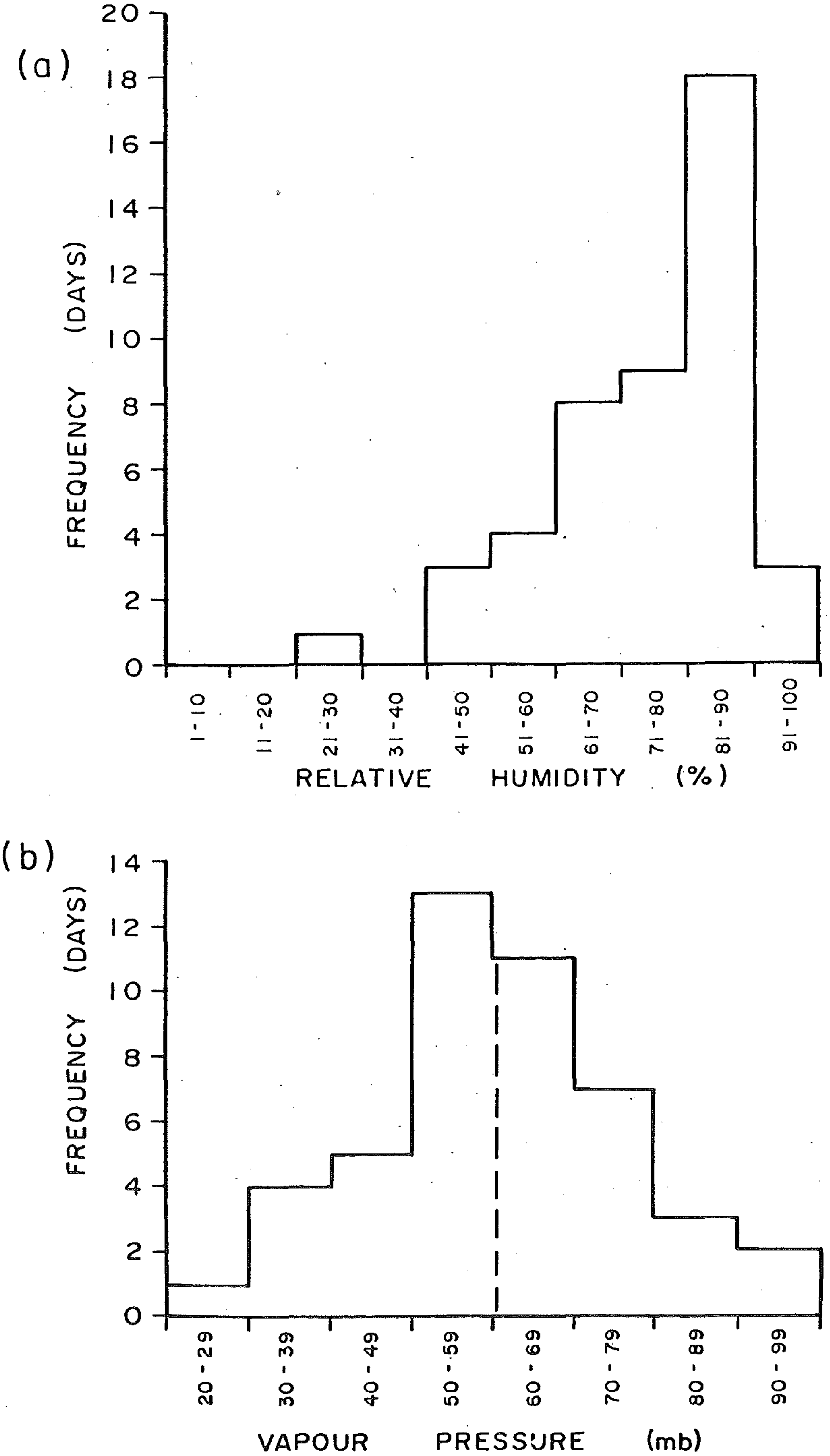


Figure 6.10a,b FREQUENCY HISTOGRAMS OF (a) RELATIVE HUMIDITY AND (b) VAPOUR PRESSURE ON DAYS OF HIGH SENSIBLE HEAT TRANSFER.

exactly half of the days recorded a mean vapour pressure of less than 6.11 mb (figure 6.10b). As noted by Diamond (1953) and Boyd (1967), for air temperatures only several degrees above freezing, relative humidities may be relatively high yet evaporation from melting snow may still take place. This simply reflects the low moisture capacity of cool air. The actual vapour pressures and temperatures for the 46 days are illustrated in figure 6.11.

Of the twenty-three days with a negative vapour pressure gradient over the snowpack, only 20 could be used in the calculation of potential evaporation, because of missing meteorological data. Ten days occurred in the May to July period and an equal number from August to October. No seasonal trend was apparent in the data with mean values for the two periods being 2.43 mm d^{-1} and 2.29 mm d^{-1} respectively. Four days were characterized by rates higher than 4.0 mm d^{-1} and a maximum of 5.70 mm d^{-1} was reached on July 30, 1975.

One should remember that the above results are only in terms of potential evaporation. Actual evaporation rates are dependent on the amount of available energy which for the sample would be expected to be largely supplied by sensible heat. The energy requirements of potential evaporation for the twenty day sample was calculated in addition to the daily input of sensible heat. On average Q_H supplied 72% of the energy required by the latent heat of vapourization, and for ten days supplied more heat than was required. If the assumption is made that inputs of Q_H were used exclusively for evaporation and any surplus was

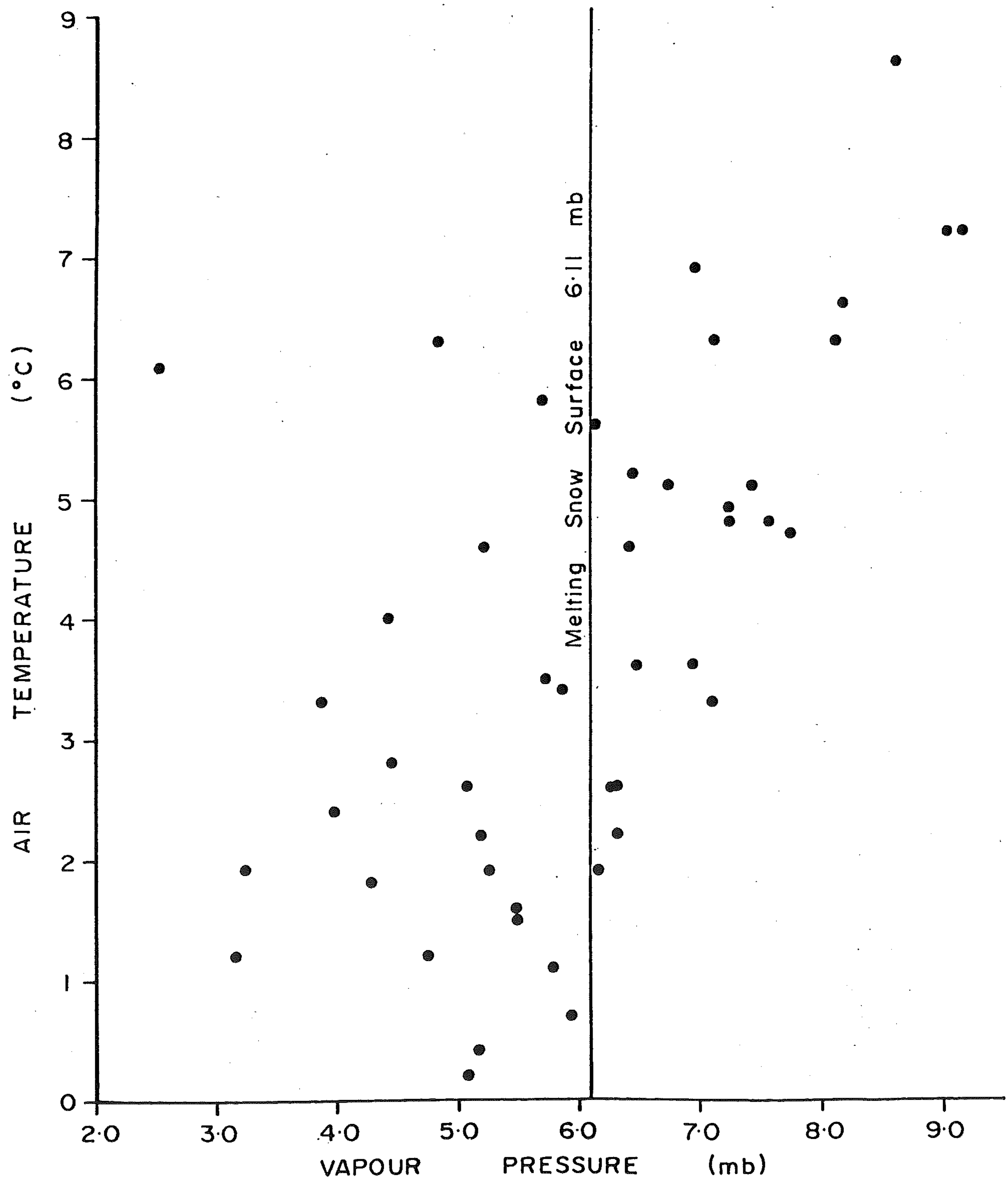


Figure 6.11 AIR TEMPERATURE AND VAPOUR PRESSURE ON DAYS OF HIGH SENSIBLE HEAT TRANSFER.

consumed in melt, evaporation rates averaged 1.28 mm d^{-1} in the May to July period and 1.53 mm d^{-1} from August to September. Five days produced rates higher than 2.00 mm d^{-1}, with a maximum of 3.28 mm d^{-1} being recorded on September 26, 1979.

For the days when Q_H was less than that required by the latent heat of vapourization, the remaining heat could be supplied by radiation (Diamond 1953).

The calculated evaporation rates are similar to many of the high values reported elsewhere (table 6.14), and are almost identical to the potential evaporation rates reported by Golding (1978) and Louie (1977) during Chinooks. However, these rates are still small in comparison to ablation rates reported for melt periods in sections 6.6 and 6.7.

Further research should be conducted on snow evaporation in the Craigieburn Mountains with a focus on the seasonal importance of evaporation to ablation. This type of study would require a more detailed monitoring of the energy balance and some complimentary volumetric sampling. Attention should also be paid to the extent of snow evaporation during wind transport. Based on theoretical estimates of sublimation from blowing snow made by Schmidt (1972), this type of evaporation in windy alpine environments probably far exceeds that occurring at the surface of the snowpack.

CHAPTER VII

CONCLUSIONS

7.1 REVIEW OF MAIN CONCLUSIONS

The snow environment of the Craigieburn Range has been addressed under three major themes: snowfall characteristics, snow metamorphism and snow melt. The data base employed in this study is considered to be the most reliable, extensive and long term record of alpine snow-climate conditions yet available in New Zealand, though even this was less extensive than would be desired. Much of the analysis was designed to interpret the snow environment in a manner usable for practical applications such as avalanche forecasting and snow hydrology.

The results pertaining to snowfall characteristics have indicated that the Craigieburn Range climate is most typical of coastal-transition or intermontane climates as found in North America. Total precipitation is less than that for maritime mountains but greater than in most interior continental regions. Snowfall contributes approximately 25 - 40% of the total annual precipitation above 1500 m and may occur at any time throughout the year, although the major accumulations are from July to September. Similarly, rain may fall in any month but is least likely in August when snow inputs are maximized.

Many winter storms occur as mixed precipitation events characterized by air temperatures which fluctuate near the freezing mark and minimum air temperatures which average only 3.5^{o}C below freezing and rarely fall below -6^{o}C. The mean elevation of the storm freezing levels in the winter months ranges from 1100 to 1300 m during periods of maximum precipitation but usually descend 400 - 500 m more in the coldest phases of the storms. Freezing levels are strongly related to wind direction; the lowest levels being associated with southerly airstreams and the warmest with the north-west. Approximately 600 m separates the mean monthly elevation of the freezing level for these wind directions during the winter months. Many snow storms are associated with shifts in wind direction from the north-west to southerly directions, especially those which occur during the passage of a typical cold front.

Most snow storms are not characterized by the high intensity, large magnitude events typical of coastal maritime locations, but again are more representative of inland-coastal conditions. Over 75% of the storms analyzed produced less than 25 mm of precipitation and a majority were 40 hours or less in duration. Snowfall intensities are usually low with 75% of the snowfall occurring at a rate of less than 2.5 mm hr^{-1}. However, there is generally one storm per year which produces more than 25 mm of snow at a rate greater than 2.5 mm hr^{-1}.

The magnitude of storm precipitation was observed to increase with elevation largely due to the longer duration of storms in the alpine versus the more sub-alpine locations. The intensity of precipitation did not, however, consistently increase with elevation and in many cases was greater at lower elevations, probably due to low elevations of the cloud base.

The largest snowfalls originated in centres of cyclonic vorticity passing in close proximity to, or directly over the South Island. Forced convection in the passage of cold fronts and orographic uplifting appear to be of secondary importance in the production of heavy snowfalls. The relative importance of warm advection along warm fronts compared to vorticity remains unclear.

Under calm conditions the density of newly deposited snow is relatively low with an observed overall mean of 130 kg m^{-3} but many observations were made of densities less than 100 kg m^{-3}, values which are usually considered more typical of cold continental climates. Atmospheric riming of crystals was found to be responsible for the production of both low and high density snow.

Snow deposited or reworked by wind, an important feature of the alpine environment, had a mean density of almost twice that of new snow deposited under calm conditions and a structure similar to intermediate stages of equi-temperature metamorphism. Characteristics of the wind deposited snow conformed with some of the definitions for hard and soft slab layers offered by other researchers, although no significant difference in density-strength relationships was found between these layers and ones formed primarily by ET metamorphism within the snowpack.

ET metamorphism occurred most rapidly in new light density snow, and produced density changes of up to 50% in one day. Large amounts of free water in the snowpack, from rain or melt events, tended to accelerate ET metamorphic changes such that the effects of ET metamorphism could not be distinguished from those due to melt-freeze metamorphism.

Indirect climatic evidence suggested that temperature-gradient metamorphism should have only a very limited effect on snowpack structure. However, on average 12% of the snowpack depth was observed to be affected by TG metamorphism and three locations in the snowpack were identified at which TG crystals regularly developed. These were: (a) a near surface TG crystal which primarily developed on south facing slopes during periods of strong negative radiation balance, (b) basal depth hoar which formed at the beginning of the winter season under shallow snowcover, and (c) TG crystals which formed above and below ice crusts. The impediment of vapour transfer and the thermal discontinuity created by the ice layers were thought to be significant in the formation of this latter TG crystal.

TG metamorphism was found to be active in a range of snow densities from 100 to 400 kg m^{-3}, and produced depth hoar strata with a mean density of 280 kg m^{-3}. Some well developed depth hoar was also observed with densities below 200 kg m^{-3}, values typical of cold continental climates.

Frequent melt and rain events resulted in the almost complete inundation of the snow column by free water. These conditions caused a rapid growth in grain size through both ET and MF metamorphism and produced significant changes in the mechanical strength of the snowpack, expecially where the flow of water was impeded by ice crusts. Rapid cooling which frequently followed melt phases refroze much of the ponded water creating extremely thick ice layers, characterized by a density and hardness similar to that of 'white' lake ice.

The overall effect of new snow inputs, wind compaction and metamorphic processes was a snowpack characterized by both rapid and frequent fluctuations in mean density. From June to October, the mean density increased from 265 to over 400 kg m^{-3}, at a rate of approximately 40 kg m^{-3} per month. As the season progressed, the density of the snowpack became much more homogenous as evinced by a decrease in the coefficient of variation calculated from the densities of the different strata. Both the magnitude and trend in mean snow density were found to be between those for maritime and continental regions of North America.

Because of the relatively warm nature of the alpine climate in the Craigieburn Range, the entire snowpack rarely has a heat deficit exceeding 4 MJ m^{-3}. Hence, even brief warming periods in the middle of winter can produce rapid melting. A study of selected winter melt periods revealed that sensible heat transfer was the major contributor to snowmelt while net radiation was negative at this time. Within each melt period, the days of greatest heat supply to the snowpack were always ones in which some rainfall occurred. Although precipitation heat flow was small, appreciable quantities of heat were received from condensation.

For the selected spring melt periods solar radiation was the single largest heat source but net radiation supplied on average only 30% of the total heat input.

Similar to the results for the winter melt periods, sensible heat transfer dominated the heat supply, comprising on average 57% of the total. The highest heat input days were again ones on which some rain fell. Precipitation heat flow never exceeded 8% but latent heat flow on these days exceeded that from net radiation. The dominance of sensible heat transfer in spring snowmelt is not typical of many continental climates where net radiation usually dominates. Large inputs of sensible heat reflect the windy nature and warm air temperatures most frequently associated with north-westerly airflows. For over 65% of the days noted to have above average temperatures and windspeeds, a north-westerly wind direction predominated. Although the mean relative humidity on these days exceeded 74%, exactly half had a net vapour pressure gradient away from the snowpack, producing an average potential evaporation of 2.5 mm d^{-1}. Considering only the heat supplied by sensible heat transfer, the rates of actual evaporation were similar to those for Chinook conditions in Canada reaching as high as 4.0 mm d^{-1}. However such losses are small relative to the melt rates which ranged from 13 - 23 mm d^{-1} and on one occasion attained 65 mm d^{-1}.

7.2 SOME IMPLICATIONS

A number of important implications can be drawn from the results of this study, one of the most important of which deals with the classification of snow regions in

New Zealand. The results of this study generally indicated that the Craigieburn Range snow environment is most typical of a coastal-transition or intermontane climate, whereby both maritime and continental effects play a role. The original classification of snow regions by the IHD Technical Subcommittee on Snow (1969) combined the Craigieburn Range with other areas such as Temple Basin into the Canterbury Foothills subregion. Temple Basin is located on the Main Divide and is very much dominated by a maritime climate, characterized by much heavier precipitation and hence snowfall than that experienced in the rainshadow region of the Craigieburn Range. Greater snowfalls alone would tend to produce a snowpack structure quite different than that which evolves under the relatively shallow snow depths of the Craigieburn Range. Observations made during this study suggest that marked differences exist in a range of snow characteristics, from elevation of freezing levels and the composition of precipitation, to the frequency and rate of melt, between the more inland mountains and those of the coastal regions. Within the west coast mountain regions of North America, snow regions are delimited across a maritime to coastal continuum on the basis of snowfall and snowpack characteristics. Although the distances are smaller in New Zealand, it would appear that some of the major differences in snow environments occur from west to east across the Southern Alps and this distinction of regions should be included in any future regional snow classification.

The results of this study also have a number of important implications for avalanche research and forecasting.

In general there are two classes of avalanches, (a) direct action avalanches, which occur during or immediately following storm cycles, and (b) climax avalanches which occur because of some inherent instability in the snowpack structure which usually develops over time. Within maritime climates where snowfalls are large and rainfall occurs regularly during the winter, direct action avalanches dominate. Forecasting of such events relies mainly on meteorological observations. In contrast, within the more continental areas where snowfalls are generally smaller and the snowpack shallower, climax avalanches dominate, often due to the formation of depth hoar. Forecasting of these events relies much more on observations of snowpack structure.

Within the Craigieburn Range, conditions were observed which favoured both direct and climax avalanching. On average, one storm per year deposited sufficient snow at a high enough rate to pose, according to overseas experience, a direct avalanche hazard. Wind is also a key factor in snow slope loading and was observed to produce both soft and hard slab strata; the latter form is known to produce some of the most dangerous avalanches overseas. The compaction of snow into slab layers is assisted by the presence of needle crystals and rime, both of which are characteristic of most snowfalls in the Craigieburn Range.

Within warm maritime climates, direct action avalanching may also occur in the winter months because of either melt or rainfall events which simply overload the slope, reduce the mechanical strength of the snowpack and/or lubricate sliding layers within the existing snow stratigraphy. The prevalence of rain and melt periods in the winter months of the Craigieburn Range is also expected to create this type of direct avalanche hazard.

The potential for climax avalanching in the Craigieburn Range is created largely by the presence of temperature gradient crystals which were observed at various depths within the snowpack, over a range of elevations and throughout the winter season. Depth hoar crystals are known to have very low mechanical strength and provide perfect sliding layers for avalanching. The probability of avalanching is likely further increased when layers of TG crystals become inundated with free water. TG metamorphism was noted to regularly occur near ice crusts, which also ponded water during melt or rain events. This situation is likely to have considerable significance for avalanching.

As the potential exists for both direct action and climax avalanches in the Craigieburn Range, avalanche forecasting must employ both meteorological and snow structure observations. La Chapelle (1966) suggests the same dual data base for avalanche forecasting in coastal transition and intermontane zones of the United States.

A number of results, especially those from the snowmelt chapter, have important hydrologic implications. Firstly, because of the relative scarcity of alpine climate stations, estimates of high elevation precipitation have frequently been based on the extrapolation of precipitation records from lowland locations in conjunction with area-discharge figures from streams whose headwaters originate in the alpine regions. The contribution of snowmelt to runoff is usually unknown and therefore ignored. As the results from this study have indicated, the greatest and most rapid snowmelt occurs during rain-on-snow events. In the cases studied, calculated snowmelt far outweighed the contribution from rainfall. If the contribution of snowmelt to discharge is not subtracted from the total discharge figures, serious errors may result in the estimation of alpine storm precipitation figures.

The rapid melt of snowcover during rain-on-snow events also has important ramifications for flood forecasting and reservoir management. In any forecast of maximum flood discharge, the contribution of snowmelt must be included. Similarly, if the benefits of snow as a water resource are to be maximized, regulation of streamflow through reservoirs must take into account the rapid depletion of snow during the spring and in particular during rain-on-snow events. This can only be realised if more information is obtained concerning the actual size of the resource and if suitable snowmelt models are employed and/or developed for New Zealand conditions.

The nature of the heat flows calculated in chapter six can provide some insight into which types of snowmelt models are most suitable. In general, there are two broad types of models: index and energy balance. Index models usually employ a single meteorological variable, commonly air temperature, which best represents energy exchanges with the snowpack. Snowmelt discharge is then primarily based on changes in the meteorological index. This type of model is ideal for New Zealand's needs because it requires only limited data input and would not require the establishment of extensive climate networks in the alpine zones. However, the application of index models has proven most successful in areas where reasonably consistent meteorological conditions prevail during snowmelt, such as found in forest environments. They have proven to be least accurate in application to open snow environments, especially during extreme or abnormal snowmelt situations. These are the precise conditions which dominate the periods of snowmelt in the Craigieburn Range.

The second type of snowmelt modelling couples the equations for energy exchange at the two snow interfaces with equations for heat transfer within the snowpack. Because the model is so physically based it can accommodate all melt conditions but is dependent on high quality input data. Several models of this type have been developed in recent years but many are still in a research-testing stage with the major goal of obtaining areal rather than point estimates of energy exchanges and hence snowmelt.

In terms of the New Zealand situation, snowcover energy balance modelling will be difficult to institute on a large scale, especially in view of the scarcity of climate stations and the variability of inter-regional alpine climates. Index models are much more desirable, the only drawback being the identification of indices which could account for the variation in snowmelt conditions. Prowse (1980b) has suggested the use of a wet bulb air temperature as a suitable index and has achieved reasonable success in modelling of a 22 day spring melt period which included both clear sky and rainfall periods. The principle behind the use of a wet bulb rather than a simple dry bulb temperature is that the former responds to fluctuations in both temperature and humidity. The index is at a maximum during warm rain-on-snow events but is much lower during low humidity, cloud-free periods when snowmelt is not as rapid. Further testing of this type of index must be conducted, preferably over a range of different New Zealand snow environments.

7.3 FUTURE RESEARCH

There remains a myriad of basic snow research which should be conducted in New Zealand, a majority of which focuses simply on obtaining a better understanding of the country's snow regions. The variety of demands for practical information concerning the snow environment makes the original research objectives of the IHD Technical Subcommittee on Snow (1969) seem somewhat limited in scope, although the study site locations originally identified by the Subcommittee make a useful framework for future classifications. Within each region, information concerning the variety of conditions under which snow is deposited, metamorphoses and melts should be gathered. The collection of this basic type of data is essential for the maximization of benefits from inter-regional applications of information from the broader body of knowledge concerning snow science and technology.

One of the surest and most efficient methods of obtaining descriptions of regional snow environments would be to identify or establish a single body or agency to conduct the necessary research. The omission of the Technical Subcommittee on Snow to do just this, is probably one of the major reasons why many of their original research objectives have not been achieved.

The New Zealand environment also offers excellent opportunities to pursue specific aspects of snow research which internationally are still in a frontier stage of investigation. Firstly, as previously noted by La Chapelle (1979), the location and orientation of the New Zealand Southern Alps are ideal for conducting basic research into the general principles of alpine meteorology. Rising abruptly from the sea and perpendicular to the path of major windflows, the Alps present a nearly perfect field laboratory for the testing of theoretical precipitation and windflow patterns over isolated mountain ranges. The accurate modelling of such patterns in New Zealand would have wide ranging benefits, not only for alpine climate classification but for such practical applications as avalanche and streamflow forecasting.

The complex nature of the Craigieburn snowpack, which exhibits characteristics from both maritime and continental climates, also presents the opportunity to examine relationships between snow strata which are not normally found in conjunction with one another in more homogenous snow environments. In particular, the influence of ice crusts on temperature gradients and vapour transfer should be more thoroughly researched with respect to temperature gradient metamorphism.

The presence of exceptionally thick and extensive ice layers also affords the possibility to conduct rigorous field tests on some of the more recently advanced theories of water transmission through stratified snowpacks. The advantages for conducting this type of investigation are further enhanced by the high frequency of winter melt and rainfall events.

An additional area of future research related to snowpack structure is the effect of large and rapid inputs of free water on the structure and mechanical strength of TG crystals. The role of high concentrations of free water on ET strata has been well researched but its effect on TG crystals seems to have been ignored. This omission may exist because depth hoar has not normally been observed in climates where frequent winter melt and rain events occur. The combination of TG crystals and free water has added significance for the Craigieburn snowpack because the supra-crust development of TG crystals coincides with the locations where the funicular states of water usually develop during melt events. The fragility of hoar crystals in association with the inter-strata lubrication created by large concentrations of free water has obvious ramifications for avalanche generation.

A final area of important future research is the quantification of the long term importance of heat flows to the energy balance of the snowpack. At present, there is insufficient data from the world's snow regions to allow a comparison of the importance of various heat flows according to major climatic and physiographic factors. Data from New Zealand could make a useful contribution to eventually achieving such a comparison.

The evaluation of long term trends in heat flows would also assist the selection of the most suitable meteorological indices for snowmelt modelling and forecasting. However, in order to properly evaluate snowpack energy balances for the different snow regions, steps should be taken towards obtaining areal averages rather than point estimates of the various heat flows. Energy balance equations have recently been coupled with digital terrain models to attempt such areal averaging. Because of the scarcity of alpine climate data and the diverse range of climates in New Zealand this type of areal modelling approach has considerable advantages.

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APPENDIX A: SNOW SURVEY EQUIPMENT AND METHODS

A.1. SNOW PIT ANALYSIS - photographic sequence

(a) pit excavation

(b) ram profile

(c) brushing to reveal layers

(d) hand hardness test

(e) temperature measurements

(f) density sampling

(g) density-weighing

(h) crystal identification

A.2. SNOW SURVEY EQUIPMENT

A.3. SNOW PIT PROFILE

(a) profile presentation

(b) grain type symbols

(c) hardness symbols

(d) water content symbols

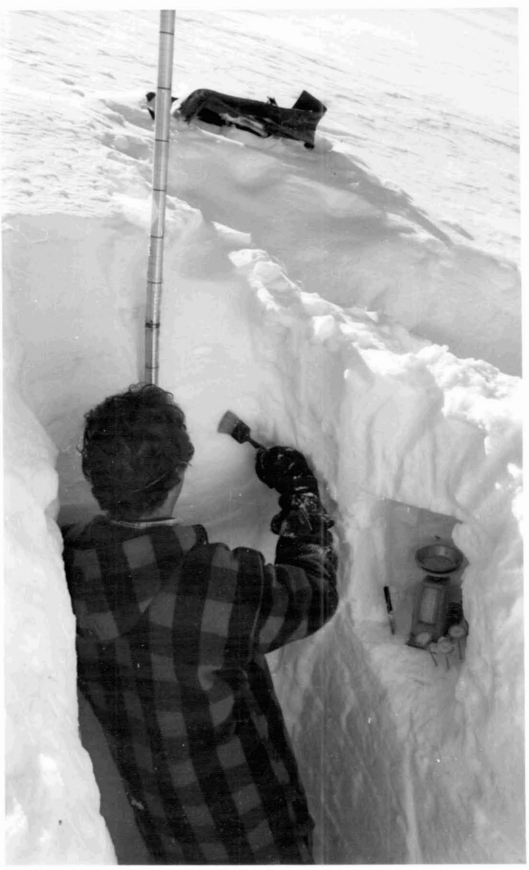
(e) ram penetrometer

A.4. SHOVEL TEST

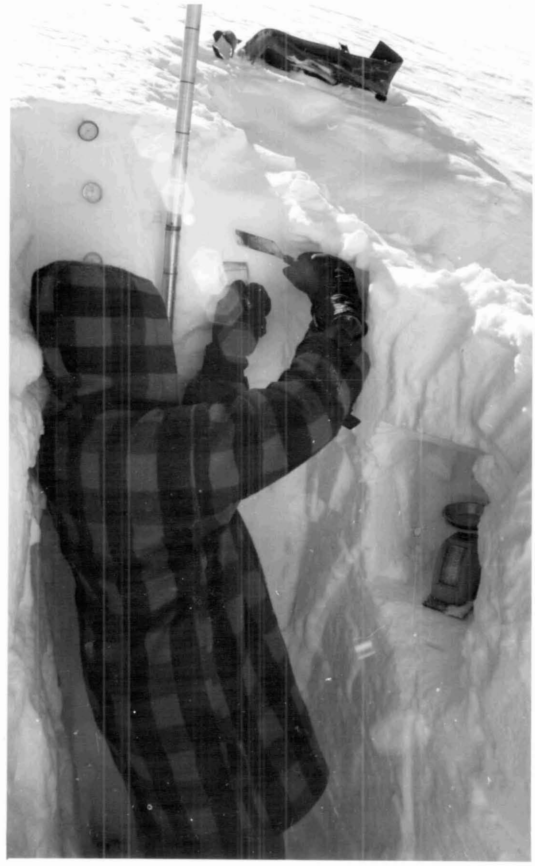
a

b

c



d

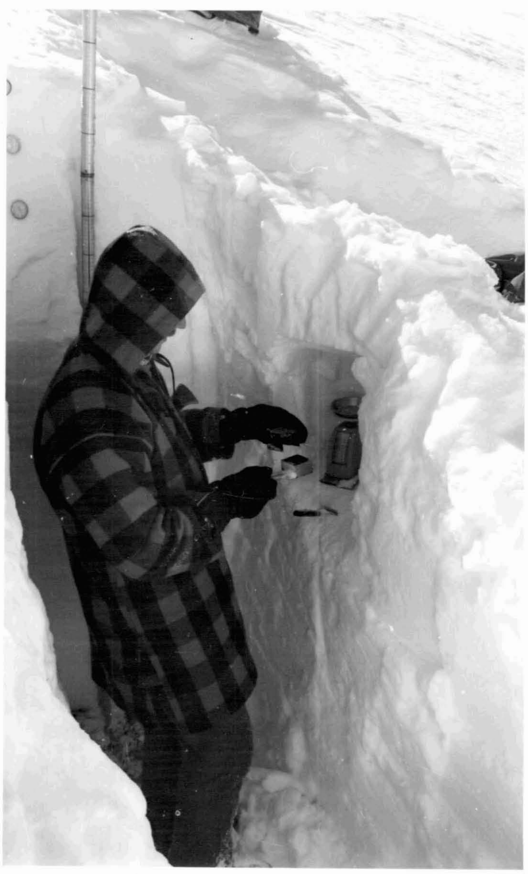


e

f

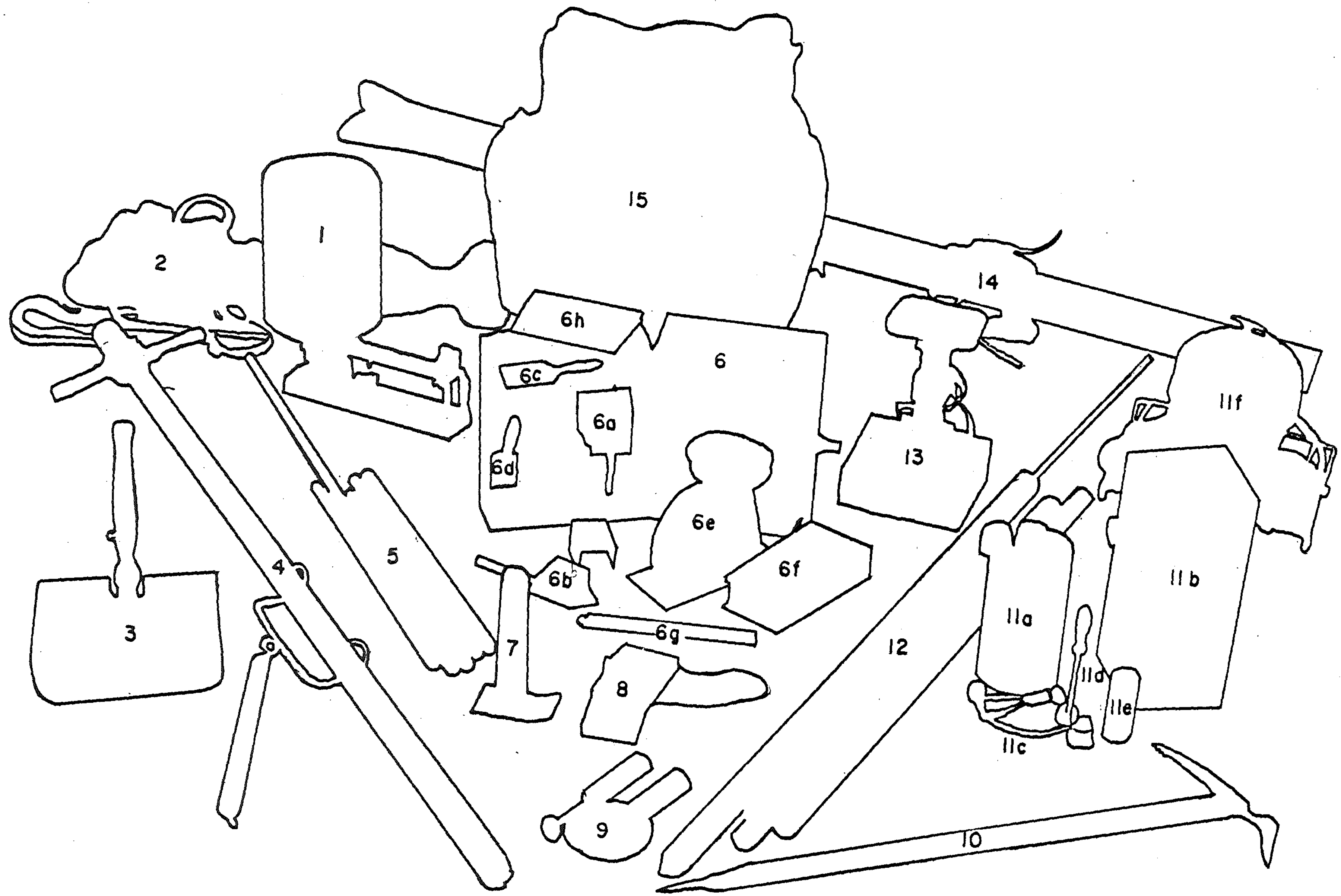
g

h

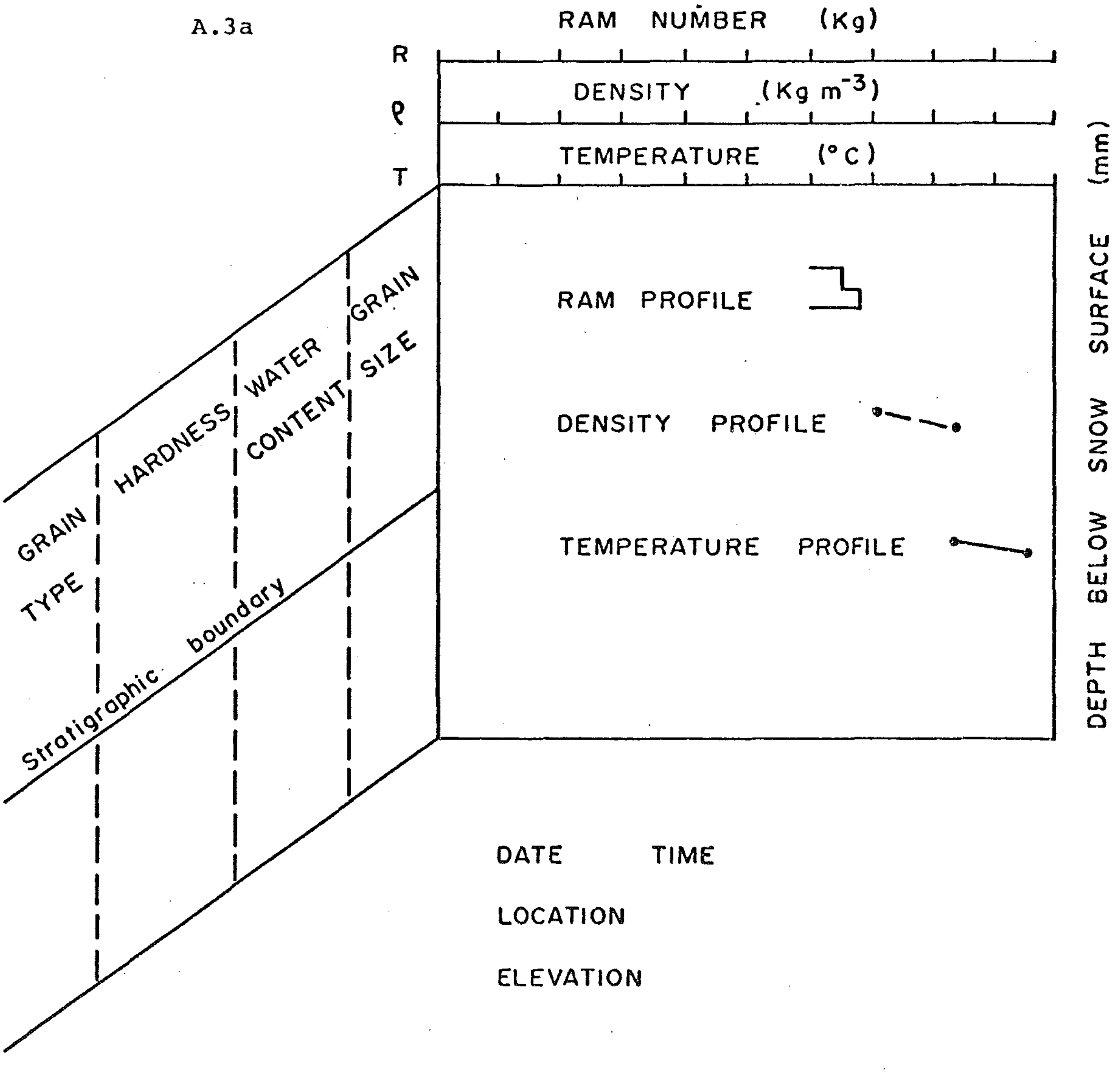


A.2 SNOW SURVEY EQUIPMENT

1. Evaporation pan-lysimeter - see section 6.
2. Avalanche cord.
3. collapsable snow shovel.
4. Federal snow density sampler.
5. depth probes - 0.5 m sections.
6. (a) 200 cc. box cutter
   (b) 100 cc. box cutter
   (c) hair brush
   (d) serrated spatula
   (e) 0 - 100 g. balance
   (f) dial stem bi-metallic thermometers
   (g) collapsable scale
   (h) waterproof field notebook
7. 8 x 30 power monocular with built in 0.1 mm graduated scale.
8. Pieps avalanche transceiver.
9. snow hardness gauges, two ranges with 0.1, 1, 10 and 100 cm^2 disks.
10. ice axe.
11. melting calorimeter
    (a) water vessel
    (b) main vessel with built-in thermometer
    (c) swing balance density scale
    (d) plug injector
    (e) 100 cc sampling tube
12. ram penetrometer
13. micro-photographic unit - light is simultaneously directed onto snow crystal from beneath (diffuse) and above (direct).
14. skis with towing bindings.
15. field pack.


15
1
2
14
6h
6
6c
6a
6d
11f
13
6e
5
4
3
6b
6f
11b
7
6g
11a
12
8
11d
11e
11c
9
10

A.3a



STANDARD SNOW PIT PROFILE USED IN THE TEXT

A.3b GRAIN TYPE

SYMBOL	DESCRIPTION
+	Freshly deposited snow initial forms can be easily recognized.
⋏	Early stages of ET metamorphism. Irregular grains, mostly rounded but often branched. Structure often feltlike.
⌀	Intermediate stages of ET metamorphism. Grains are round, often elongated and usually less than 0.5 mm in diameter; sizes are decreasing.
●	Advanced stages of ET metamorphism. Grains similar in shape to the above but larger, usually less than 2.0 mm; sizes are increasing.
o	Grains formed by MF metamorphism and usually greater than 1 mm. Shape of grains not necessarily rounded because of extensive bonding produced by refreezing of free water.

□	Early stages of TG metamorphism. Grains are angular with some flat sides on faces.
∧	Intermediate stages of TG metamorphism. Most grains in ice matrix are angular with some early growth of beaker crystals, usually between 0.5 - 1.0 mm in diameter.
△	Advanced stages of TG metamorphism. Depth hoar crystals dominate and usually greater than 1 mm in diameter.
∿∿∿	Sun or rain crust usually less than 10 mm thick.
//////	Ice layers and lens thicker than sun and rain crusts and usually more dense.
∨	Surface hoar crystals.

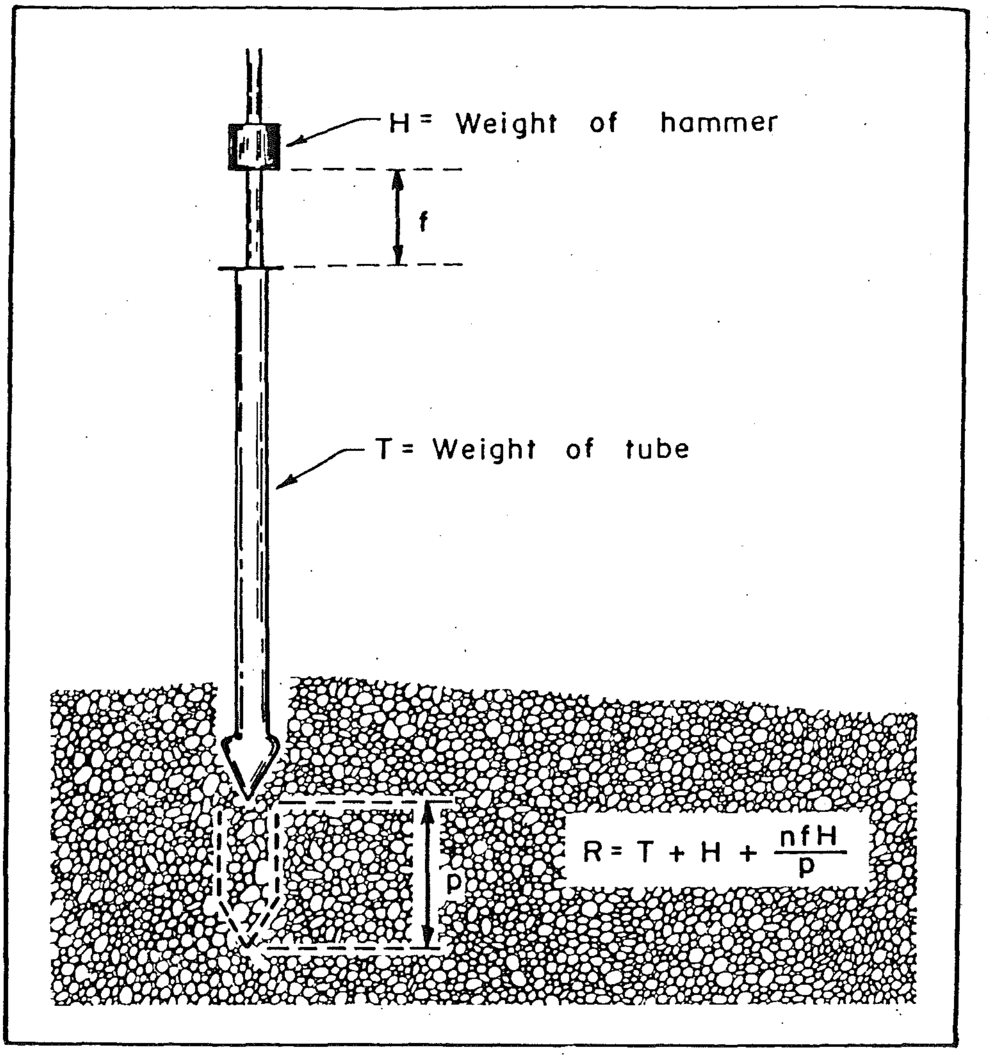
A.3c HARDNESS

SYMBOL RANGE	HAND TEST	RAM NUMBER (kg)
VSS	Fist	0 - 2
VS		
MSS	Four fingers	3 - 15
MS		
M	One finger	26 - 50
MH		
MHH	Pencil	51 - 100
H		
HH	Knife	Over 100
VH		
VHH	Thick Ice	-

In the hand test the specified object is pushed into the snow with a force of about 5 kg. A gloved hand is used in the first three levels. The symbol range refers to the ease by which the object enters the snow.

A.3d FREE WATER CONTENT

SYMBOL		
0	DRY	Snow usually but not necessarily below 0°C. Grains have little tendency to stick together in a snowball when lightly pressed in gloved hand.
1	MOIST	Snow at 0°C. No water visible even with hand lens. Snow makes good snowball.
2	WET	Snow at 0°C. Water visible as a meniscus between grains but cannot be pressed out by moderate squeezing in the hand.
3	VERY WET	Snow at 0°C. Water can be pressed out by moderate squeezing in the hand. There is still an appreciable amount of air confined within the snow.
4	SLUSH	Snow at 0°C. Snow flooded with water and containing relatively small amounts of air.



A.3e RAM PENETROMETER

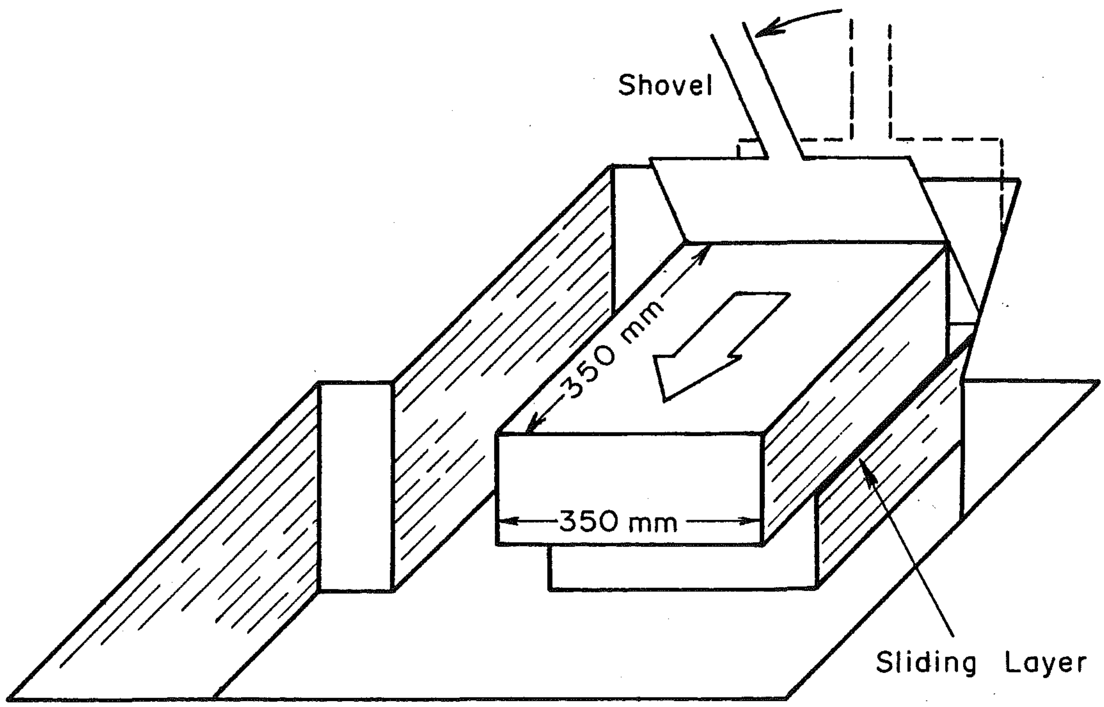
H = weight of hammer (kg)

f = fall height (cm)

T = weight of tube (kg)

p = depth of penetration (cm)

n = number of hits

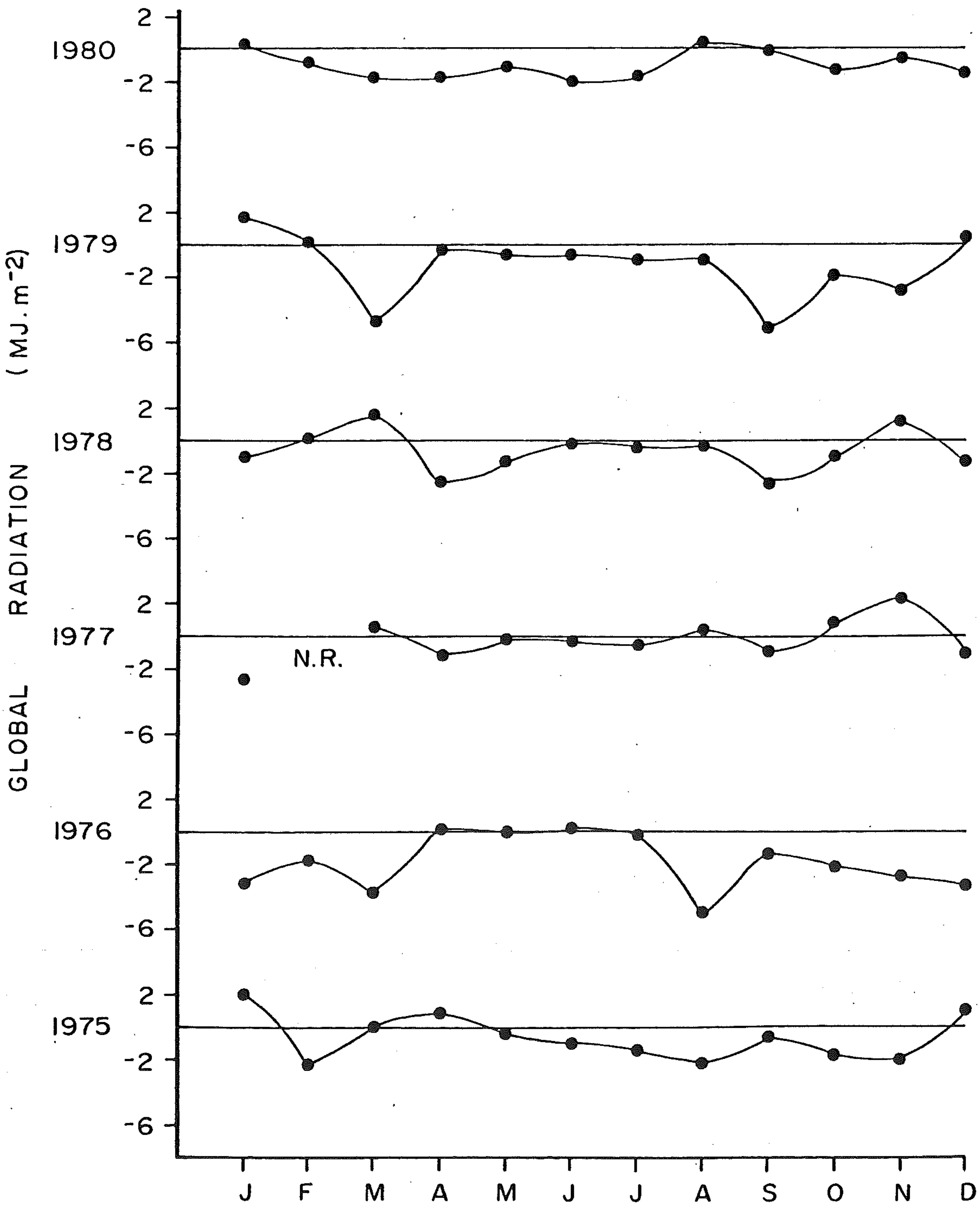


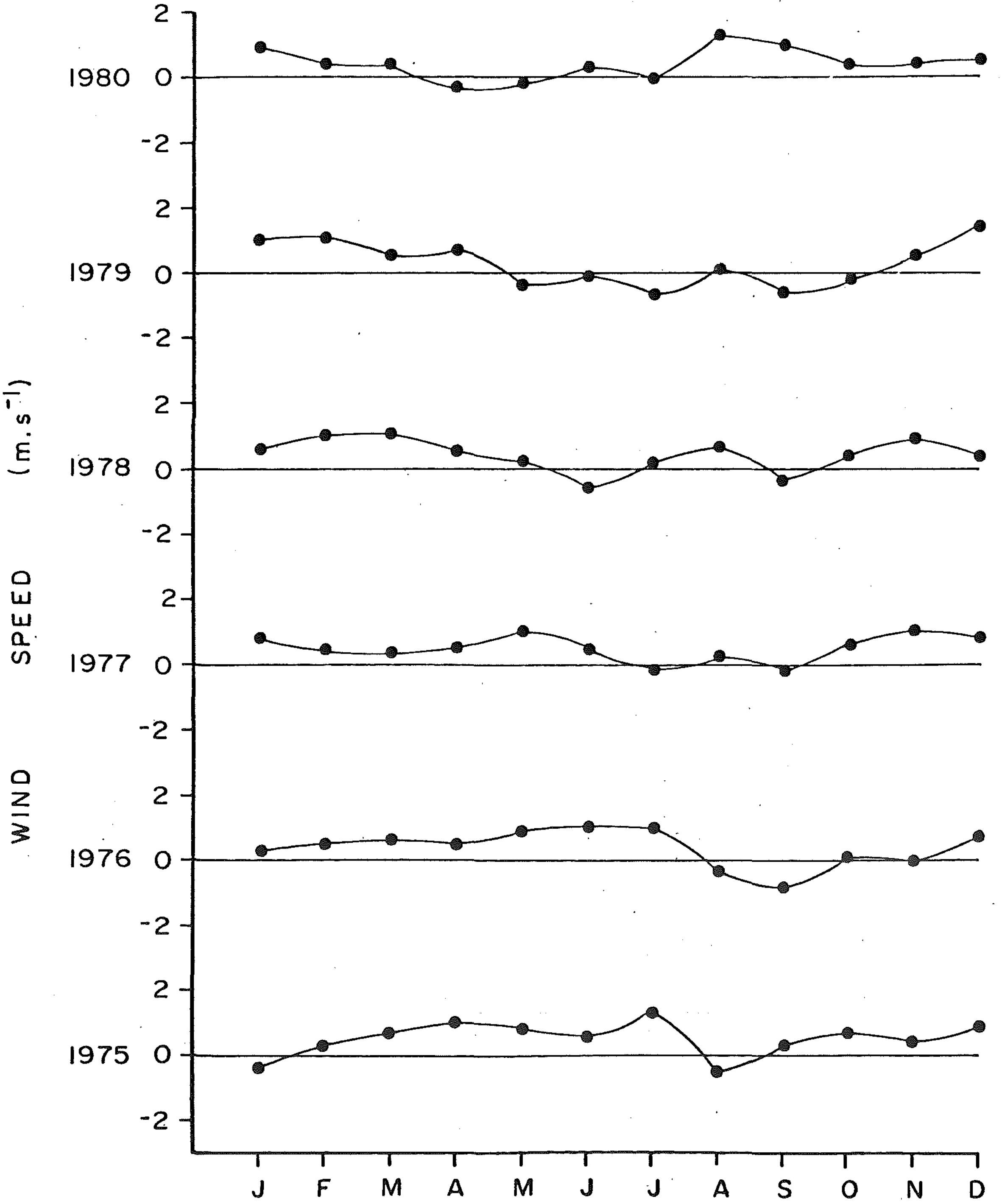
A.4. SHOVEL TEST

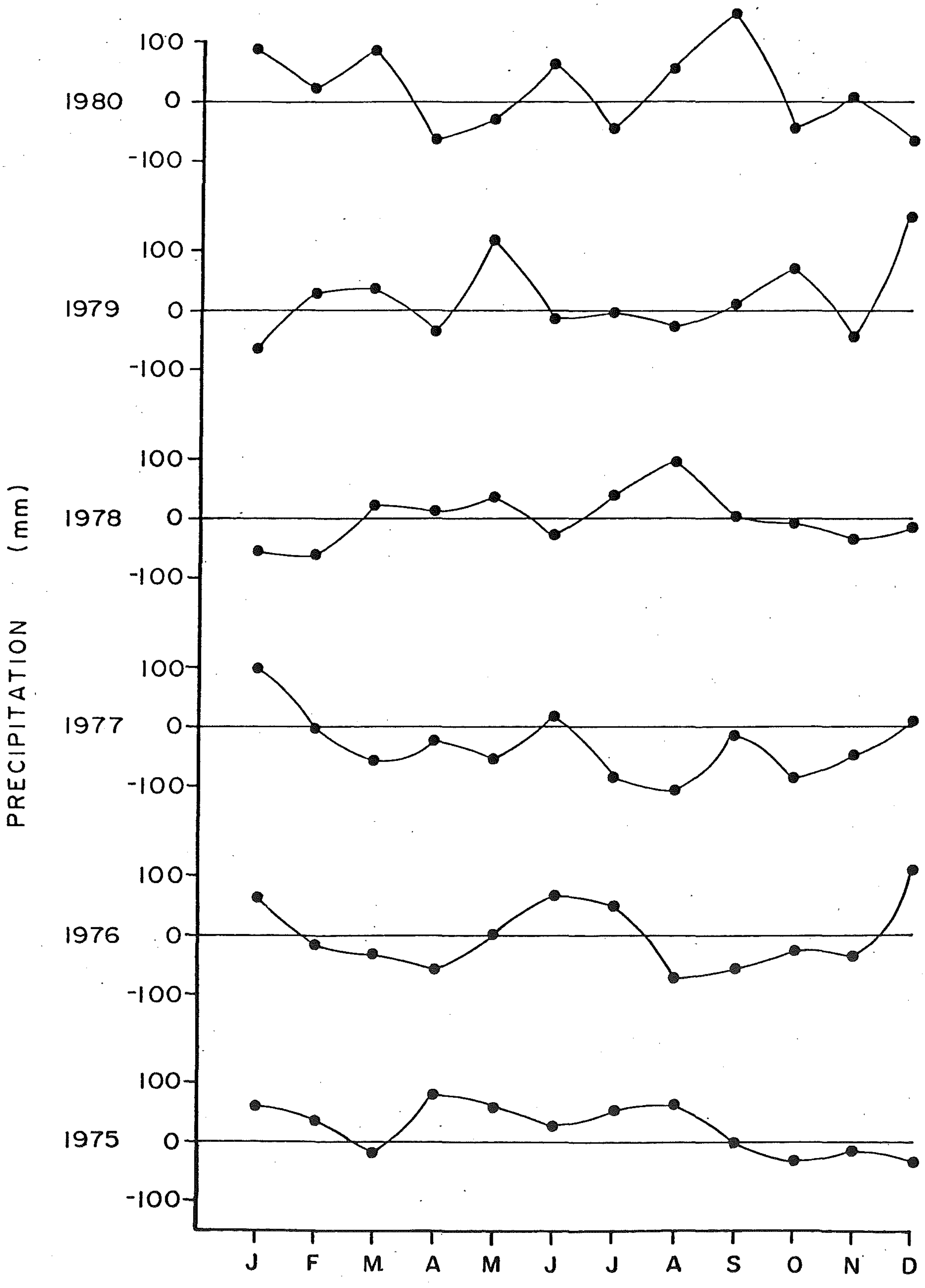
In many snow pits a shovel test was employed to help identify potential sliding layers. The technique involves excavating a vertical three sided column of snow in the face of a snow pit. The column dimensions are approximately 350 mm wide by 350 mm deep. A shovel is gradually inserted at the rear of the column and levered forward until the column fails. The crystal structure of the upper and lower faces of the sliding layer is then examined.

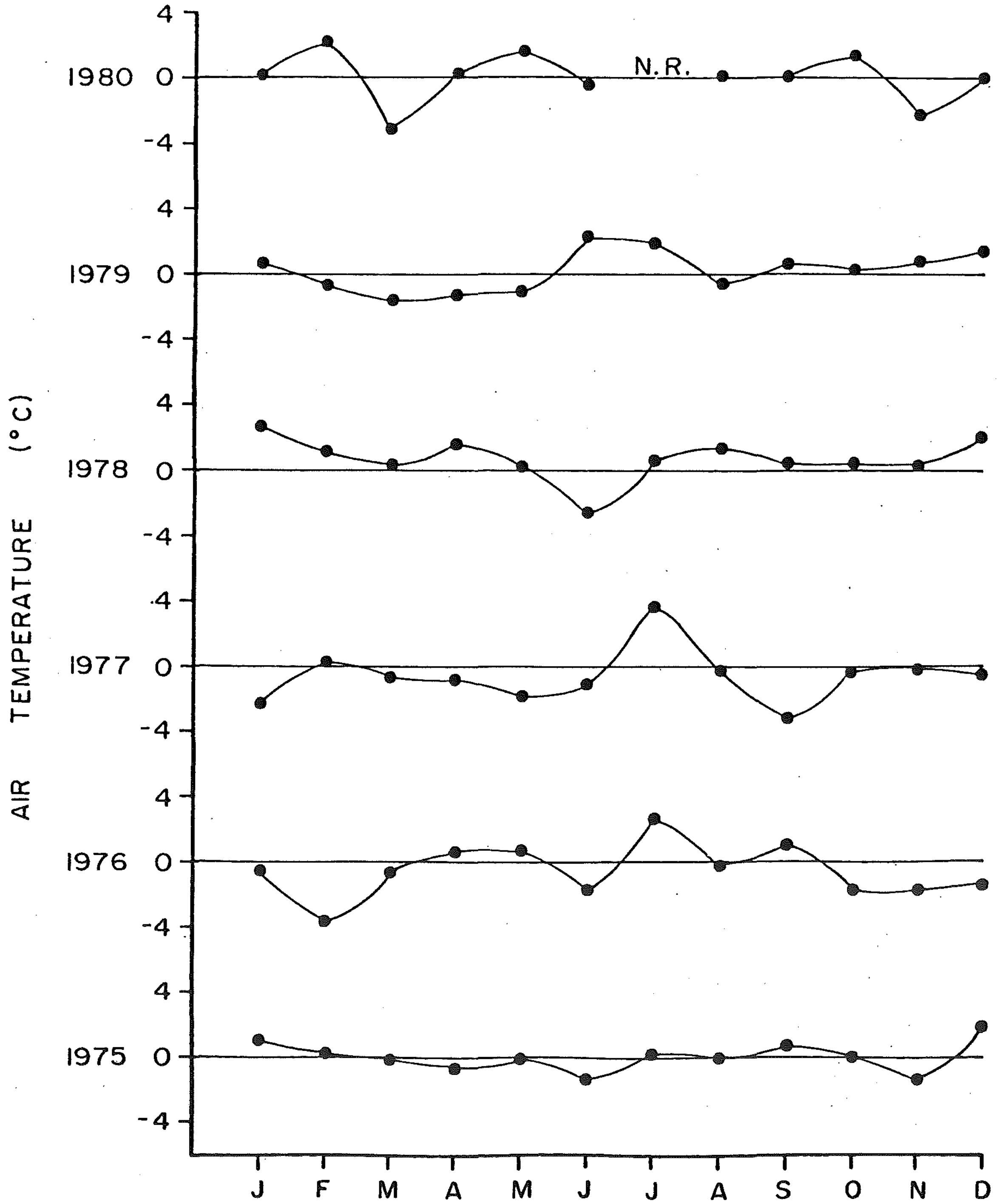
APPENDIX B MONTHLY CLIMATE RECORDS FOR 1975-1980 COMPARED TO LONG TERM MEANS.

Long term means of air temperature (5 - 12 year record), rainfall (6 - 14 year record) and windspeed are from the SB climate station; global radiation values (15 year record) are from the CF climate station.


GLOBAL RADIATION (MJ. m-2)
1980
1979
1978
1977
1976
1975
2
-2
-6
N.R.
J
F
M
A
M
J
J
A
S
O
N
D


WIND SPEED (m.s^{-1})
1980
1979
1978
1977
1976
1975
2
0
-2
J
F
M
A
M
J
J
A
S
O
N
D


PRECIPITATION (mm)
1980
1979
1978
1977
1976
1975
100
0
-100
J F M A M J J A S O N D



4
1980 0
-4
N.R.
4
1979 0
-4
4
1978 0
-4
4
1977 0
-4
4
1976 0
-4
4
1975 0
-4
AIR TEMPERATURE (°C)
J F M A M J J A S O N D

APPENDIX C MONTHLY AND YEARLY PRECIPITATION TOTALS.

C.1. MONTHLY PRECIPITATION TOTALS BY TYPE OF PRECIPITATION

(a) 1975-1976
(b) 1976-1977
(c) 1977-1978
(d) 1978-1979
(e) 1979-1980

(years span period March-February)

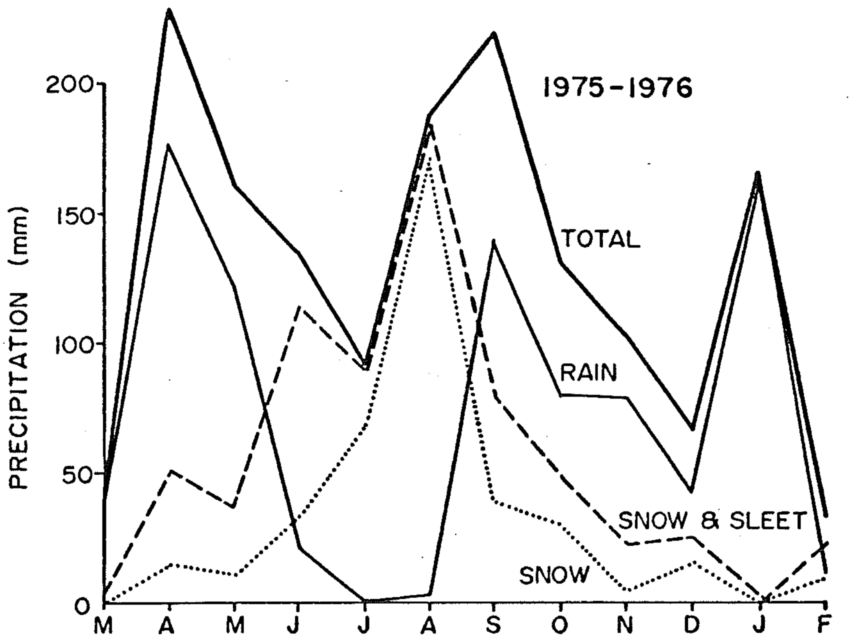
C.2. MONTHLY PRECIPITALS BY YEAR

(a) Snow
(b) Total precipitation
(c) Rain
(d) Snow plus sleet

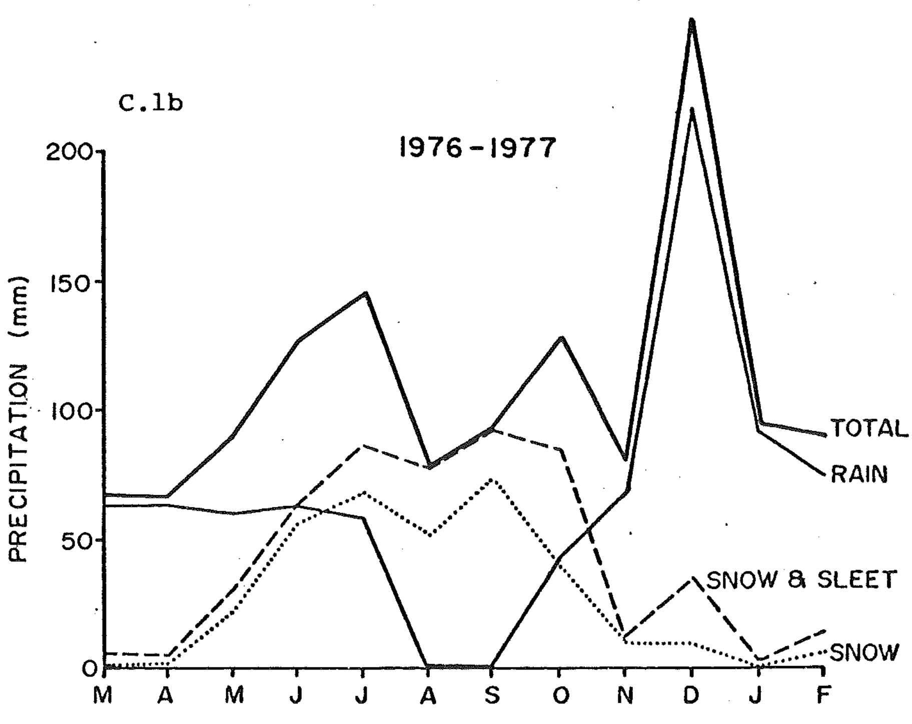
C.3. MEAN, MINIMUM AND MAXIMUM MONTHLY PRECIPITATION TOTALS, 1975-1980.

(a) Snow
(b) Sleet
(c) Rain
(d) Snow plus sleet
(e) Total precipitation

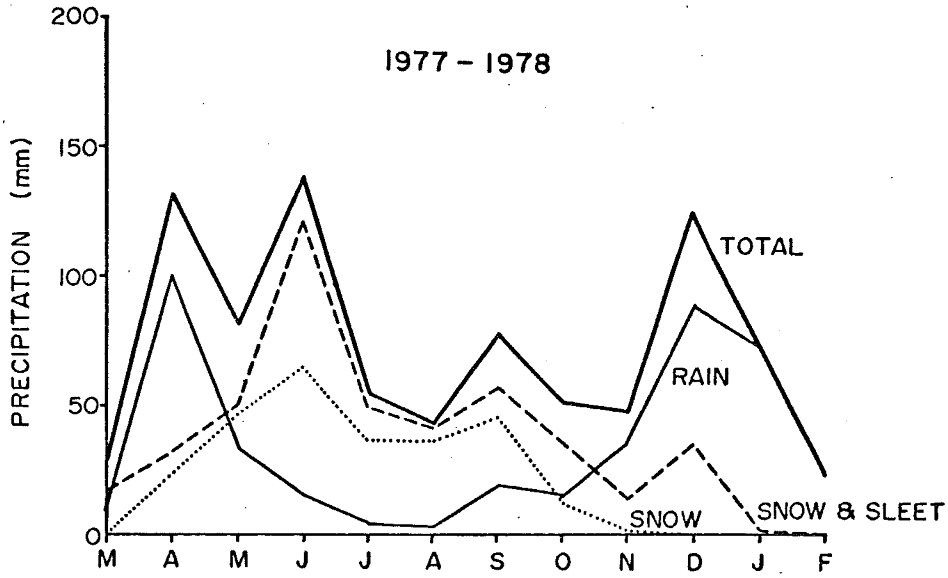
C.1a


1975-1976
200
150
100
50
0
PRECIPITATION (mm)
TOTAL
RAIN
SNOW & SLEET
SNOW
M
A
M
J
J
A
S
O
N
D
J
F

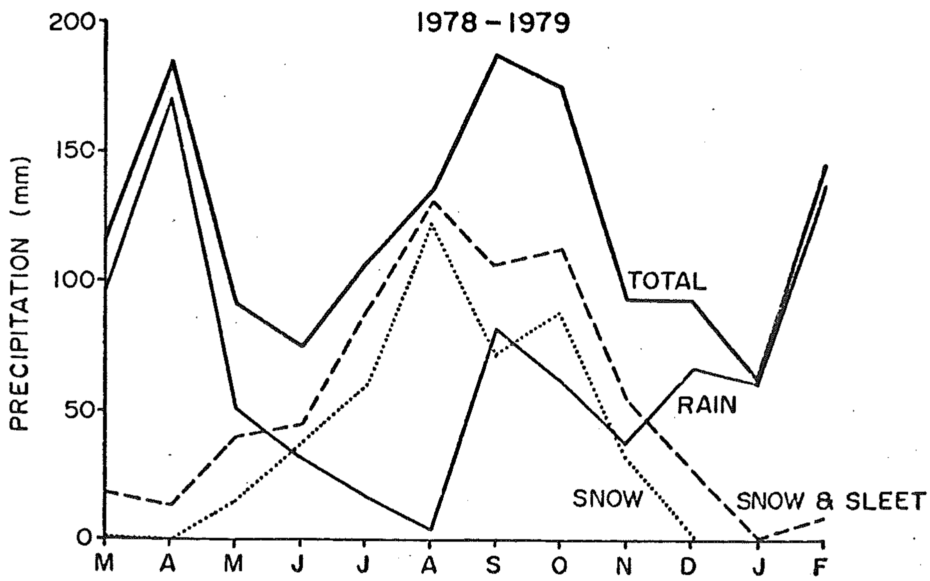
C.1b

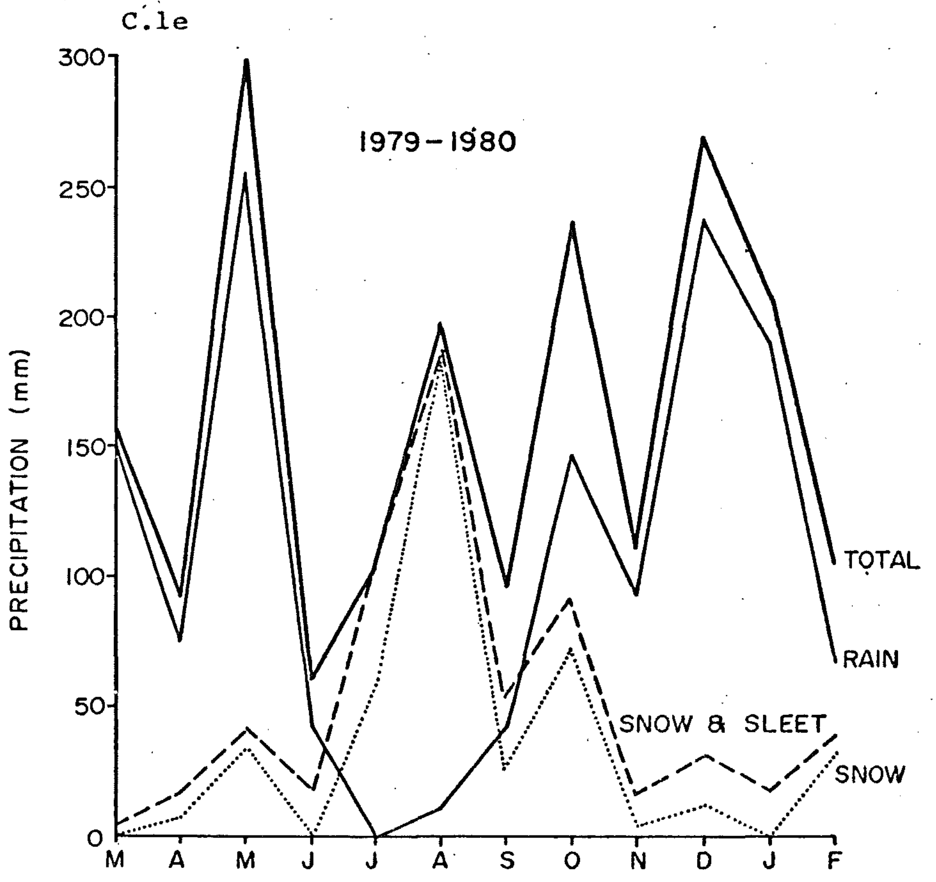

1976-1977
200
150
100
50
0
PRECIPITATION (mm)
TOTAL
RAIN
SNOW & SLEET
SNOW
M
A
M
J
J
A
S
O
N
D
J
F

C.1c

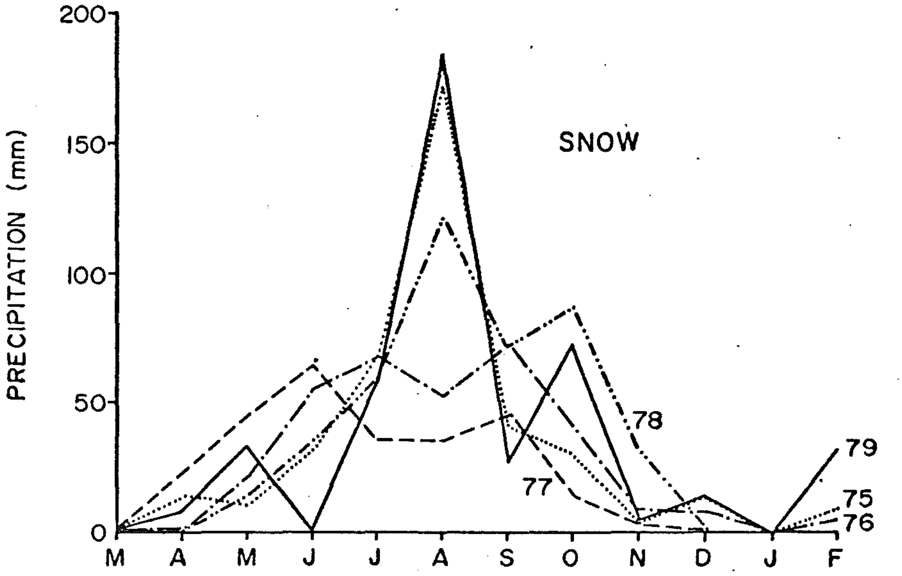

1977 - 1978
PRECIPITATION (mm)
200
150
100
50
0
TOTAL
RAIN
SNOW
SNOW & SLEET
M
A
M
J
J
A
S
O
N
D
J
F

C.1d

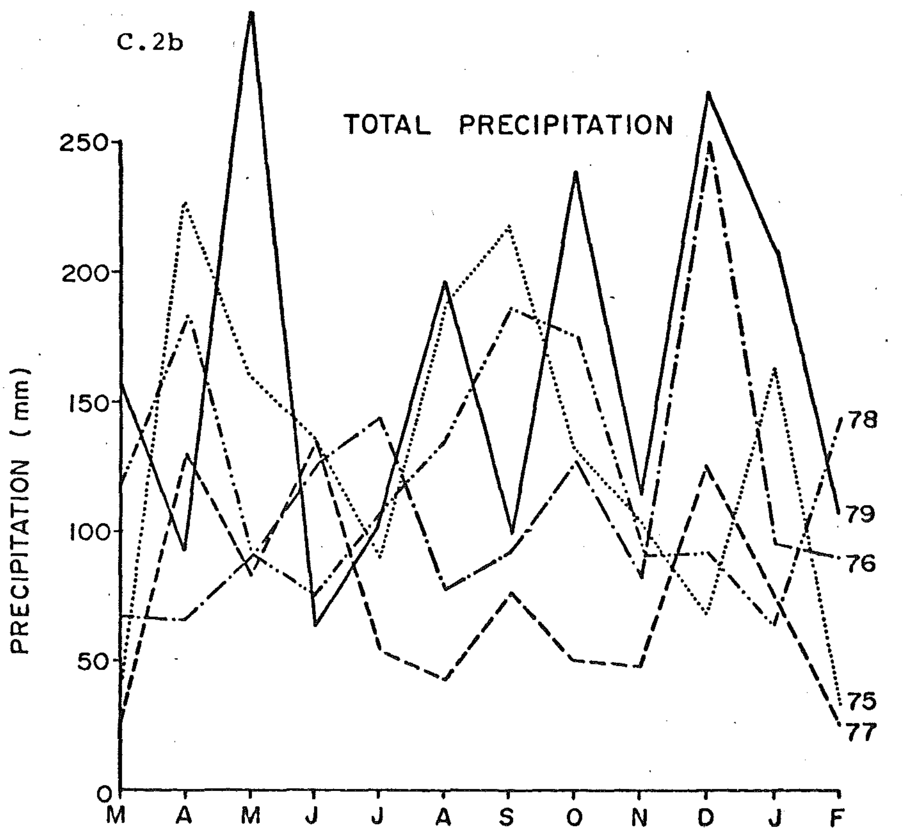

1978 - 1979
PRECIPITATION (mm)
200
150
100
50
0
TOTAL
RAIN
SNOW
SNOW & SLEET
M
A
M
J
J
A
S
O
N
D
J
F


C.le
1979–1980
PRECIPITATION (mm)
300
250
200
150
100
50
0
M
A
M
J
J
A
S
O
N
D
J
F
TOTAL
RAIN
SNOW & SLEET
SNOW

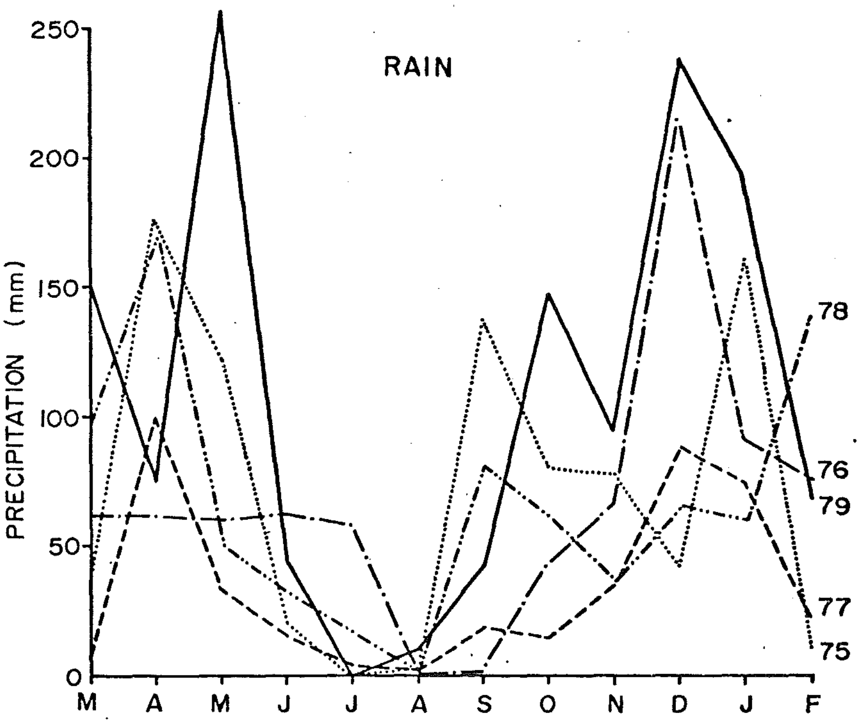
C.2a


SNOW
PRECIPITATION (mm)
200
150
100
50
0
M A M J J A S O N D J F
78
77
79
75
76

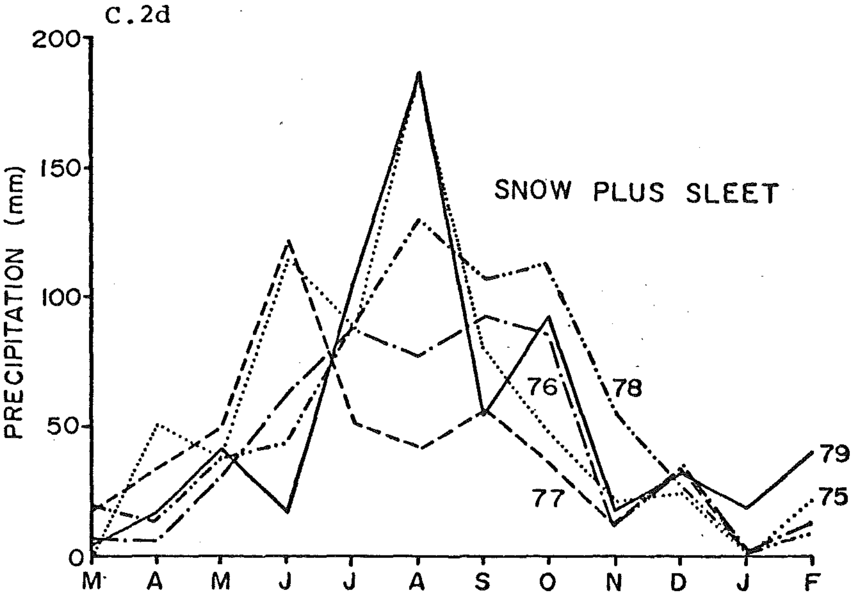
C.2b


TOTAL PRECIPITATION
PRECIPITATION (mm)
250
200
150
100
50
0
M A M J J A S O N D J F
78
79
76
75
77

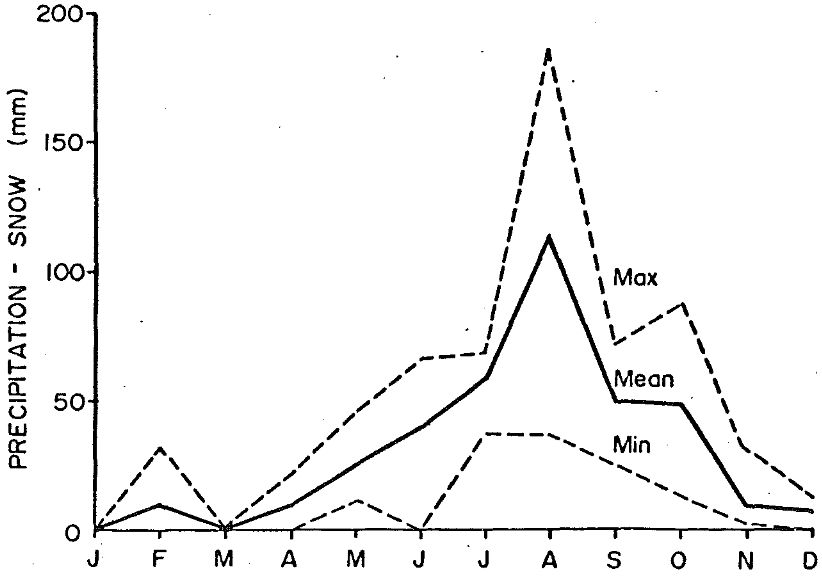
C.2c


250
200
150
100
50
0
PRECIPITATION (mm)
RAIN
M A M J J A S O N D J F
78
76
79
77
75

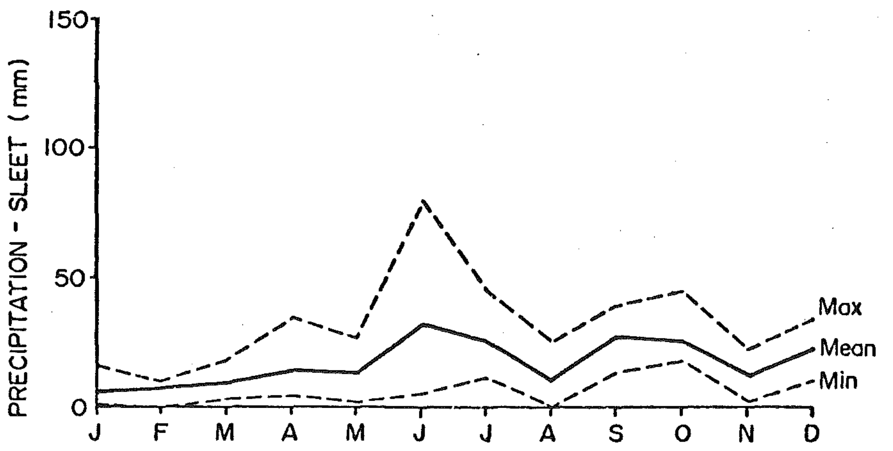
C.2d


200
150
100
50
0
PRECIPITATION (mm)
SNOW PLUS SLEET
76
78
77
79
75
M A M J J A S O N D J F

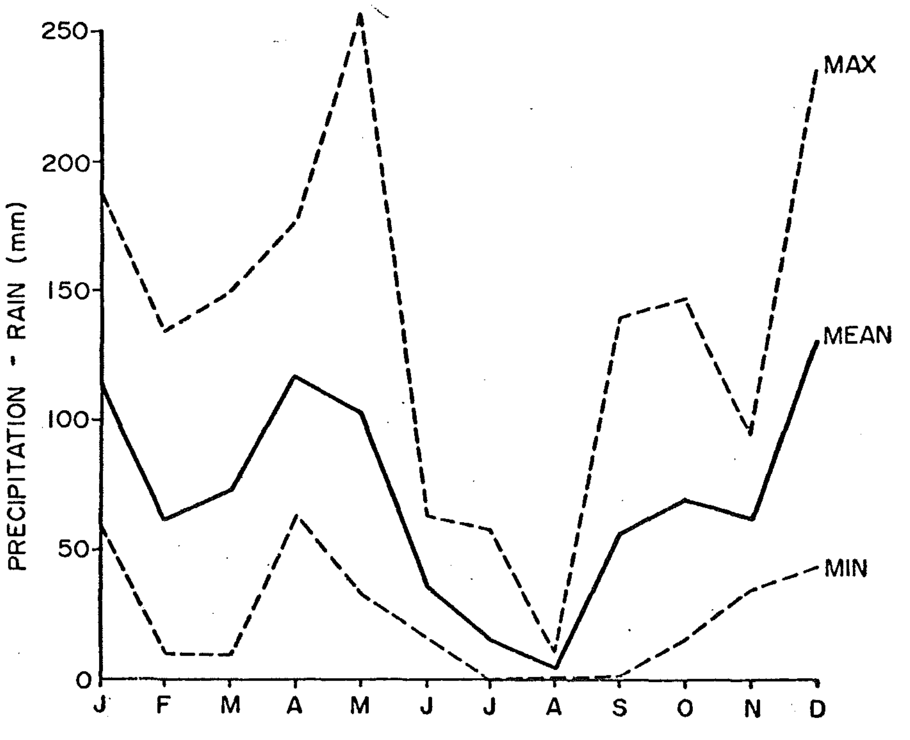
C.3a


200
150
100
50
0
PRECIPITATION - SNOW (mm)
J F M A M J J A S O N D
Max
Mean
Min

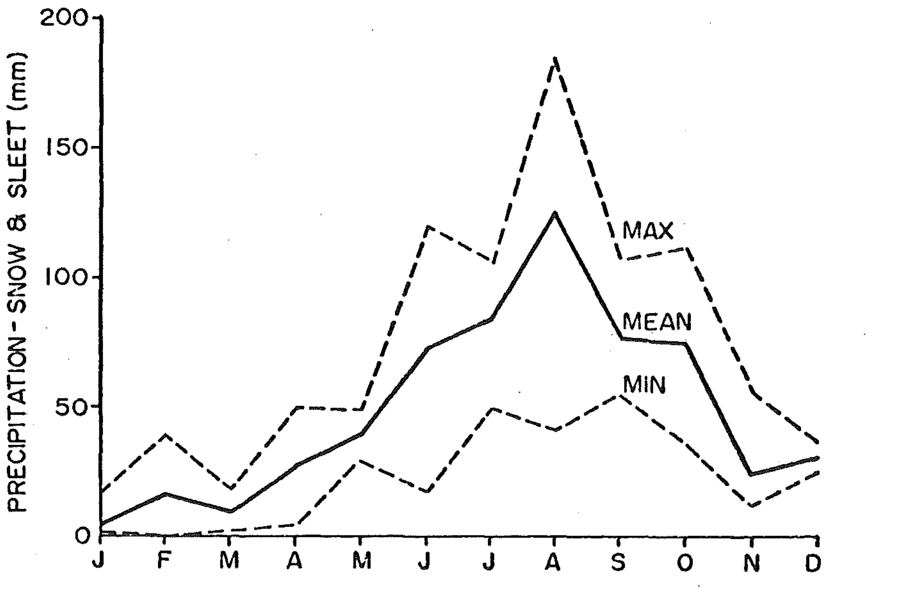
C.3b


150
100
50
0
PRECIPITATION - SLEET (mm)
J F M A M J J A S O N D
Max
Mean
Min

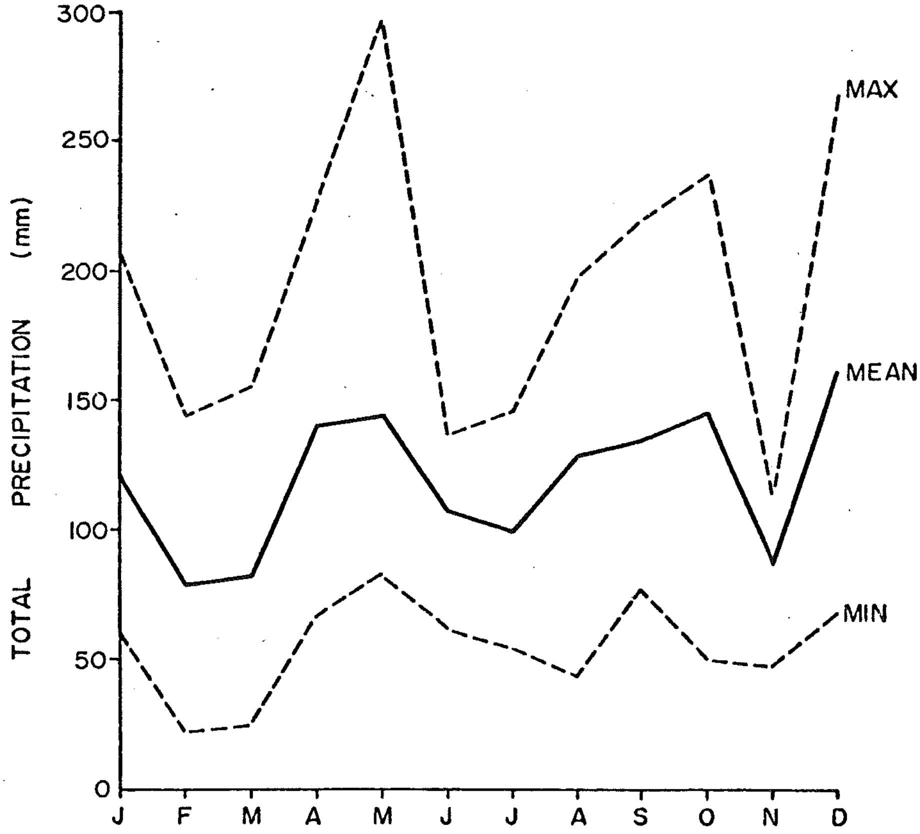
C.3c


250
200
150
100
50
0
PRECIPITATION - RAIN (mm)
MAX
MEAN
MIN
J F M A M J J A S O N D

C.3d


200
150
100
50
0
PRECIPITATION - SNOW & SLEET (mm)
MAX
MEAN
MIN
J F M A M J J A S O N D

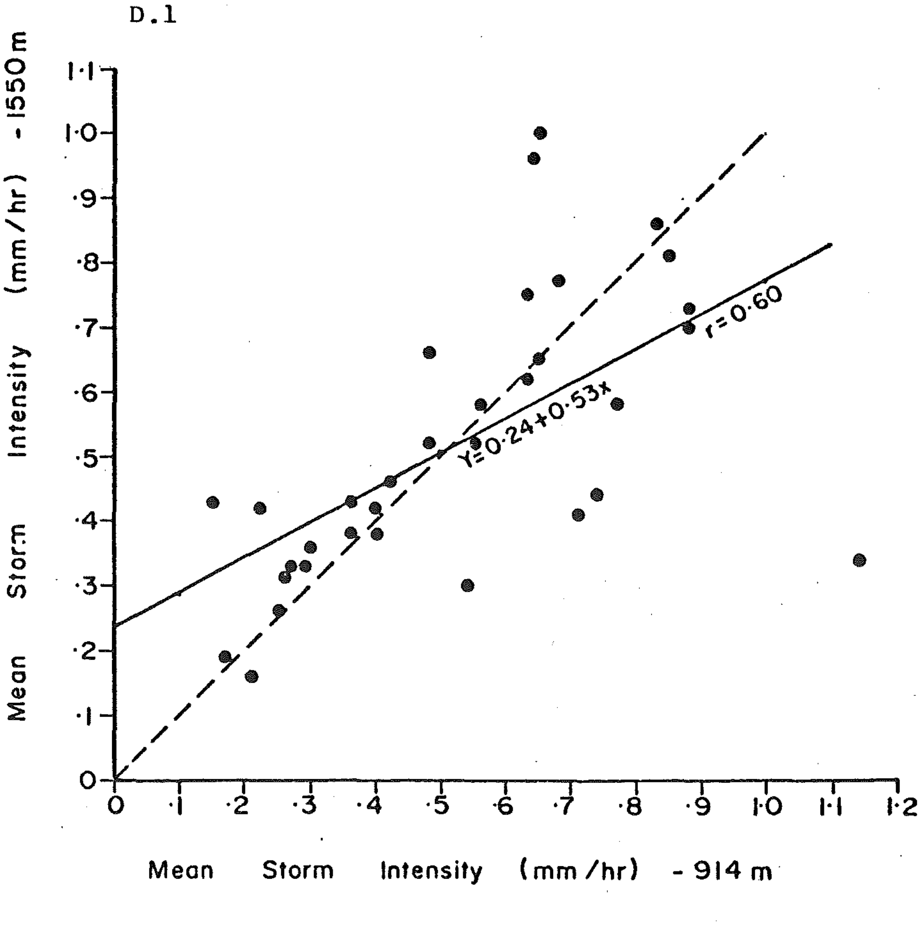
C.3e


300
250
200
150
100
50
0
TOTAL PRECIPITATION (mm)
J
F
M
A
M
J
J
A
S
O
N
D
MAX
MEAN
MIN

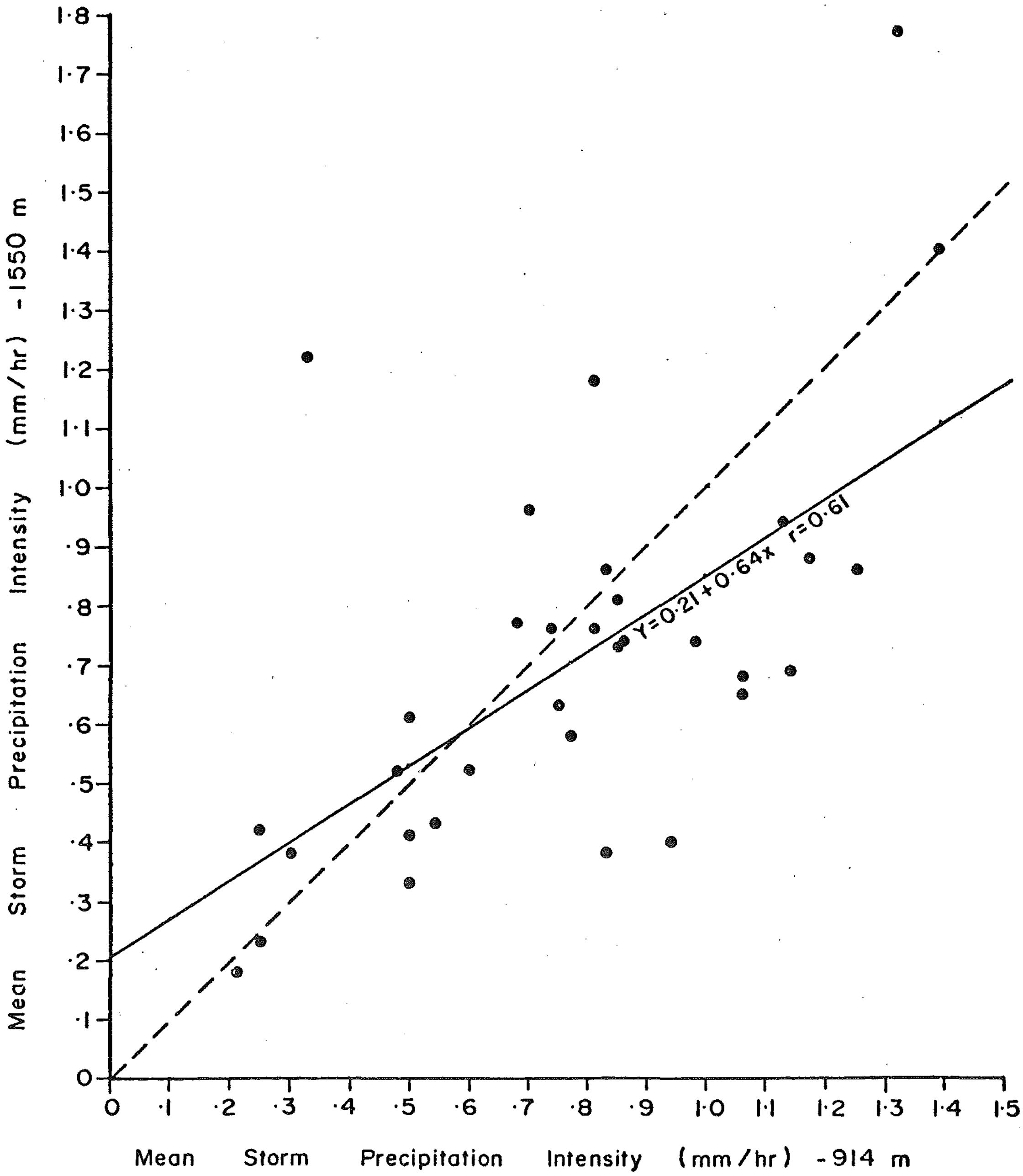
APPENDIX D PRECIPITATION CHARACTERISTICS BETWEEN THE SB AND CF CLIMATE STATIONS.

NOTE: DATA WERE STANDARDIZED FOR REGRESSION ANALYSIS.

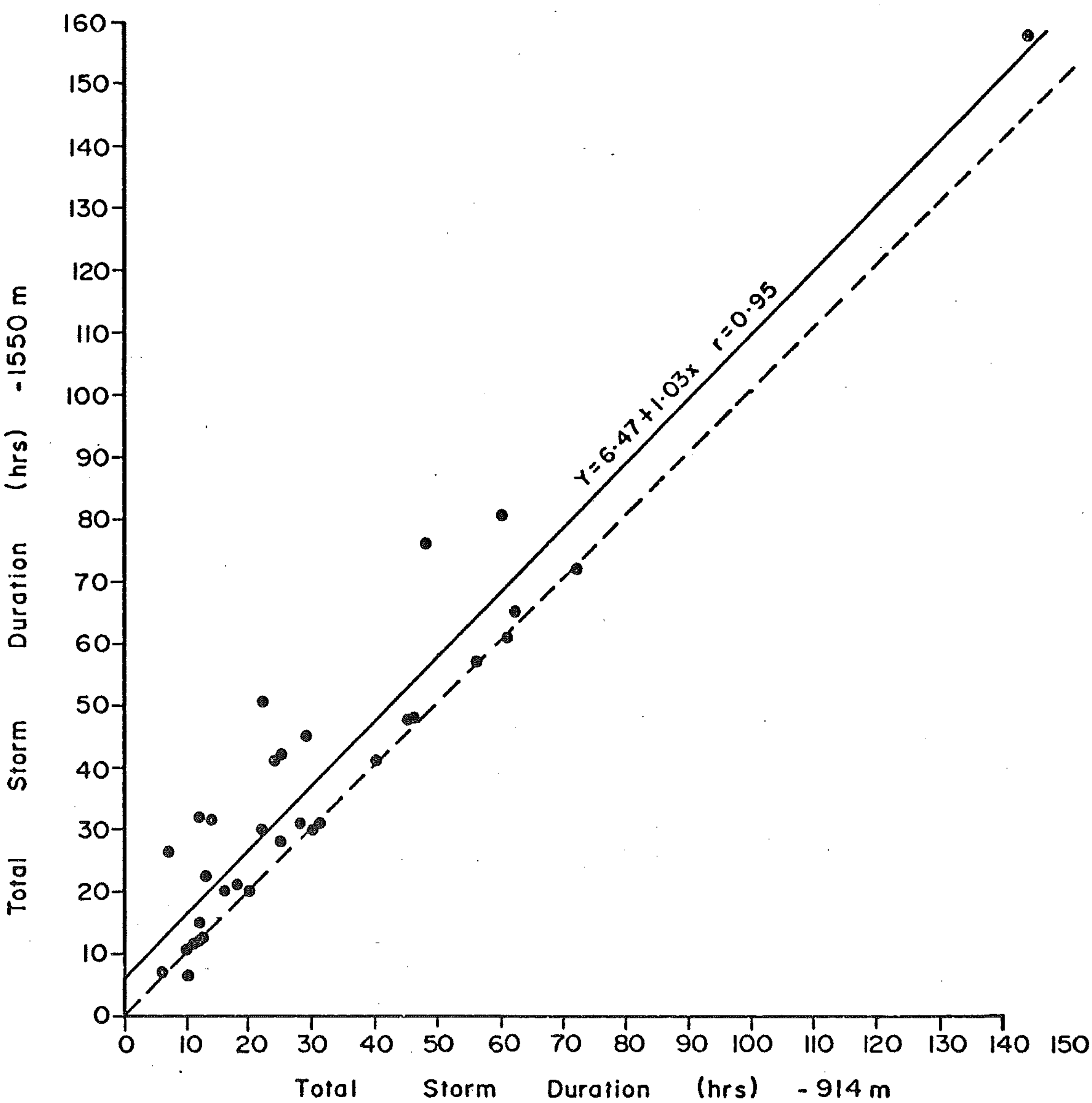
D.1


Mean Storm Intensity (mm/hr) - 1550 m
Mean Storm Intensity (mm/hr) - 914 m
Y=0·24+0·53x
r=0·60
0
·1
·2
·3
·4
·5
·6
·7
·8
·9
1·0
1·1
1·2

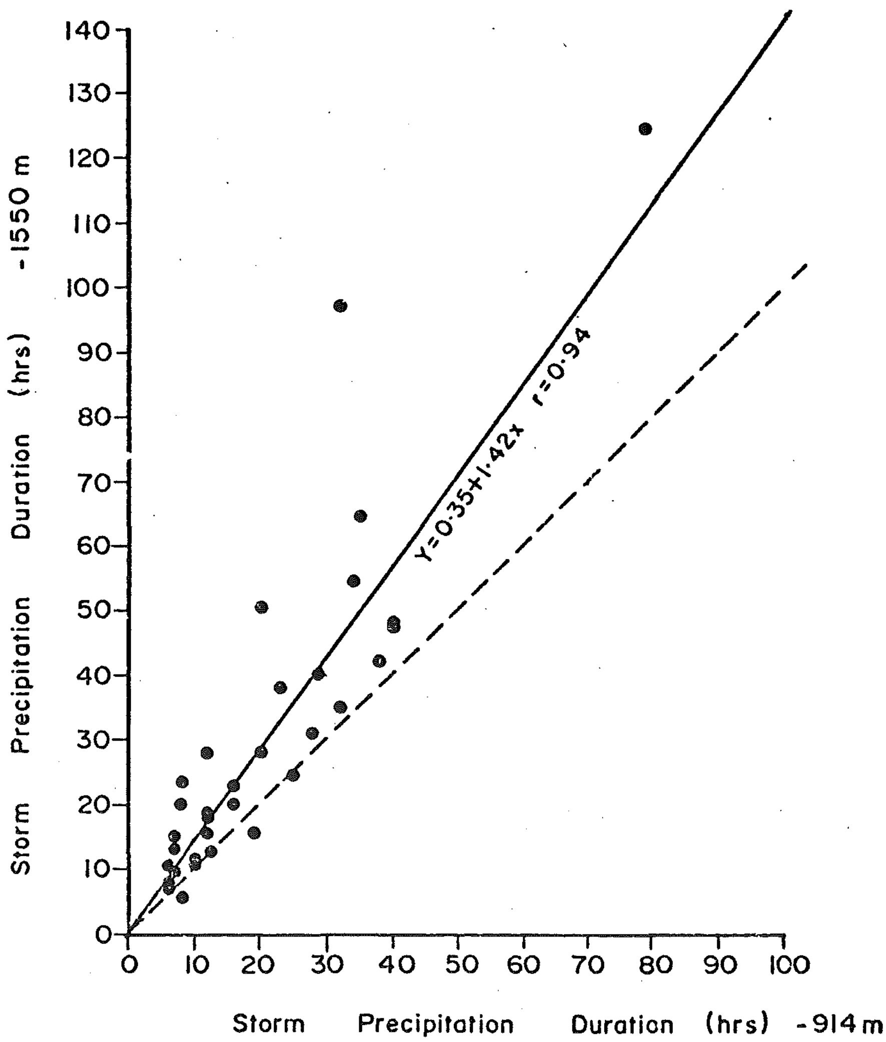
D.2


Y=0·21+0·64x r=0·61
Mean Storm Precipitation Intensity (mm/hr) -1550 m
Mean Storm Precipitation Intensity (mm/hr) -914 m
0
·1
·2
·3
·4
·5
·6
·7
·8
·9
1·0
1·1
1·2
1·3
1·4
1·5
1·6
1·7
1·8

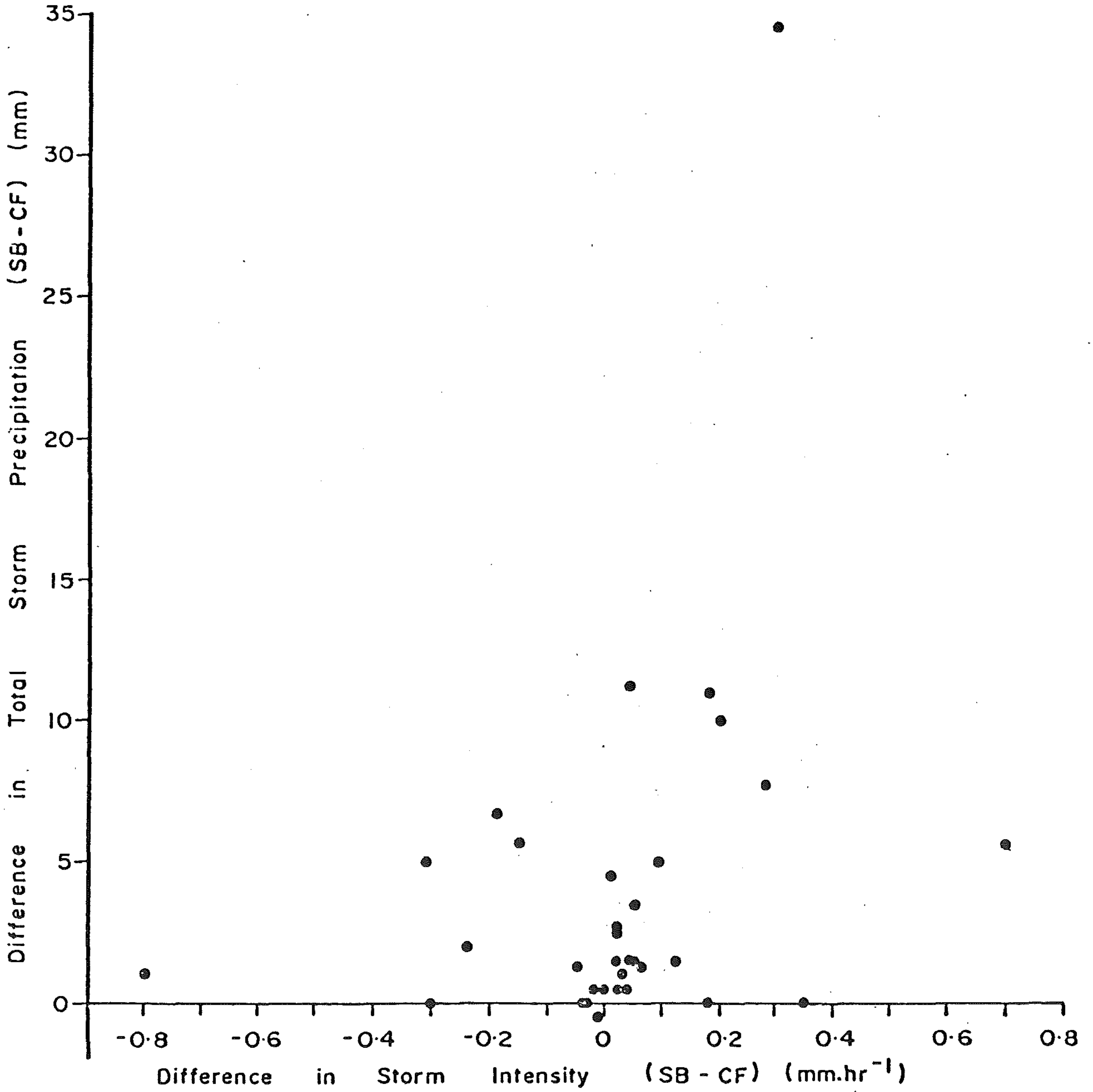
D.3


160
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
Total Storm Duration (hrs) -1550 m
Y=6·47+1·03x r=0·95
0 10 20 30 40 50 60 70 80 90 100 110 120 130 140 150
Total Storm Duration (hrs) - 914 m

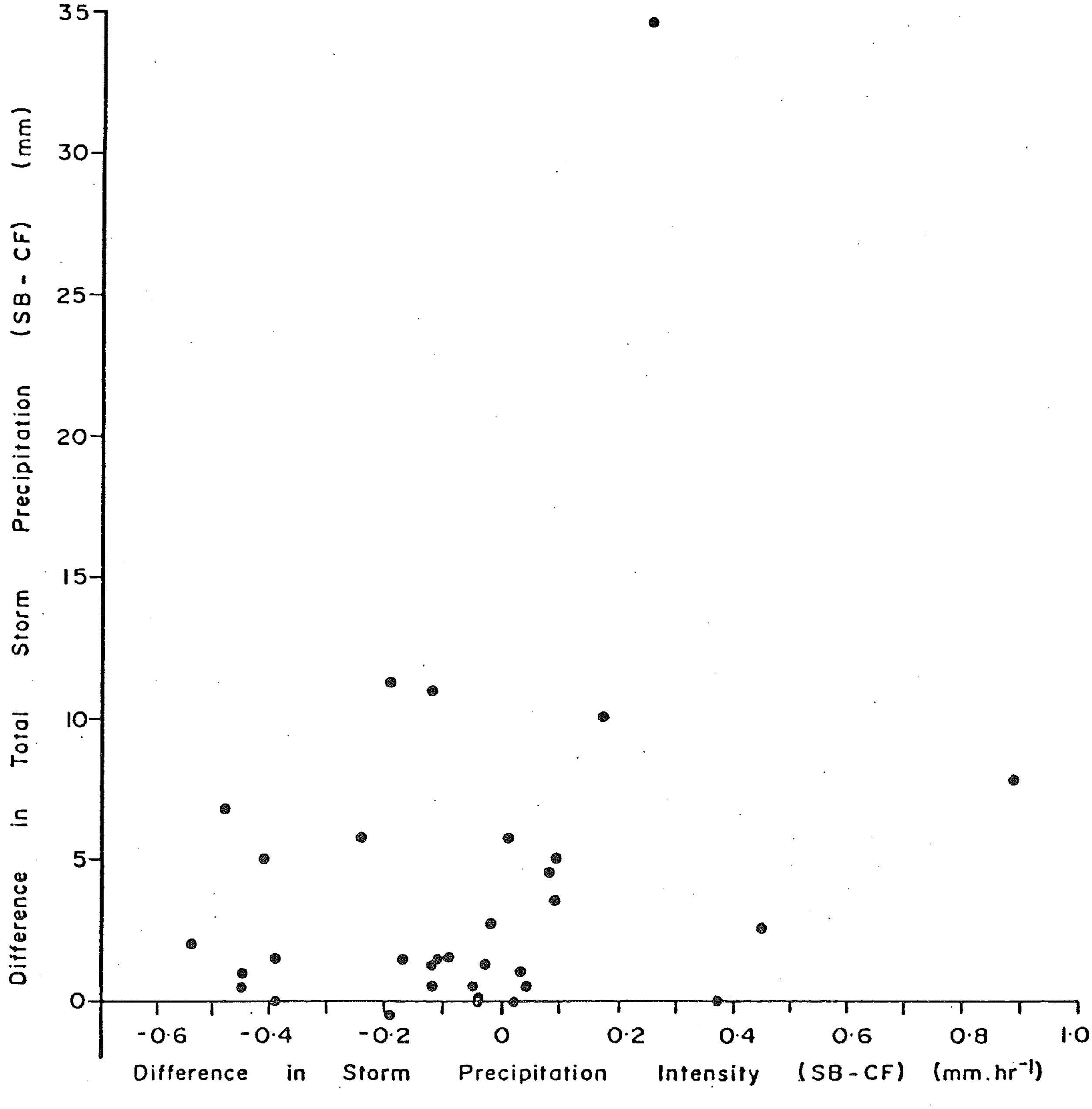
D.4


Storm Precipitation Duration (hrs) -1550 m
Storm Precipitation Duration (hrs) -914 m
Y=0·35+1·42x r=0·94
0
10
20
30
40
50
60
70
80
90
100
110
120
130
140

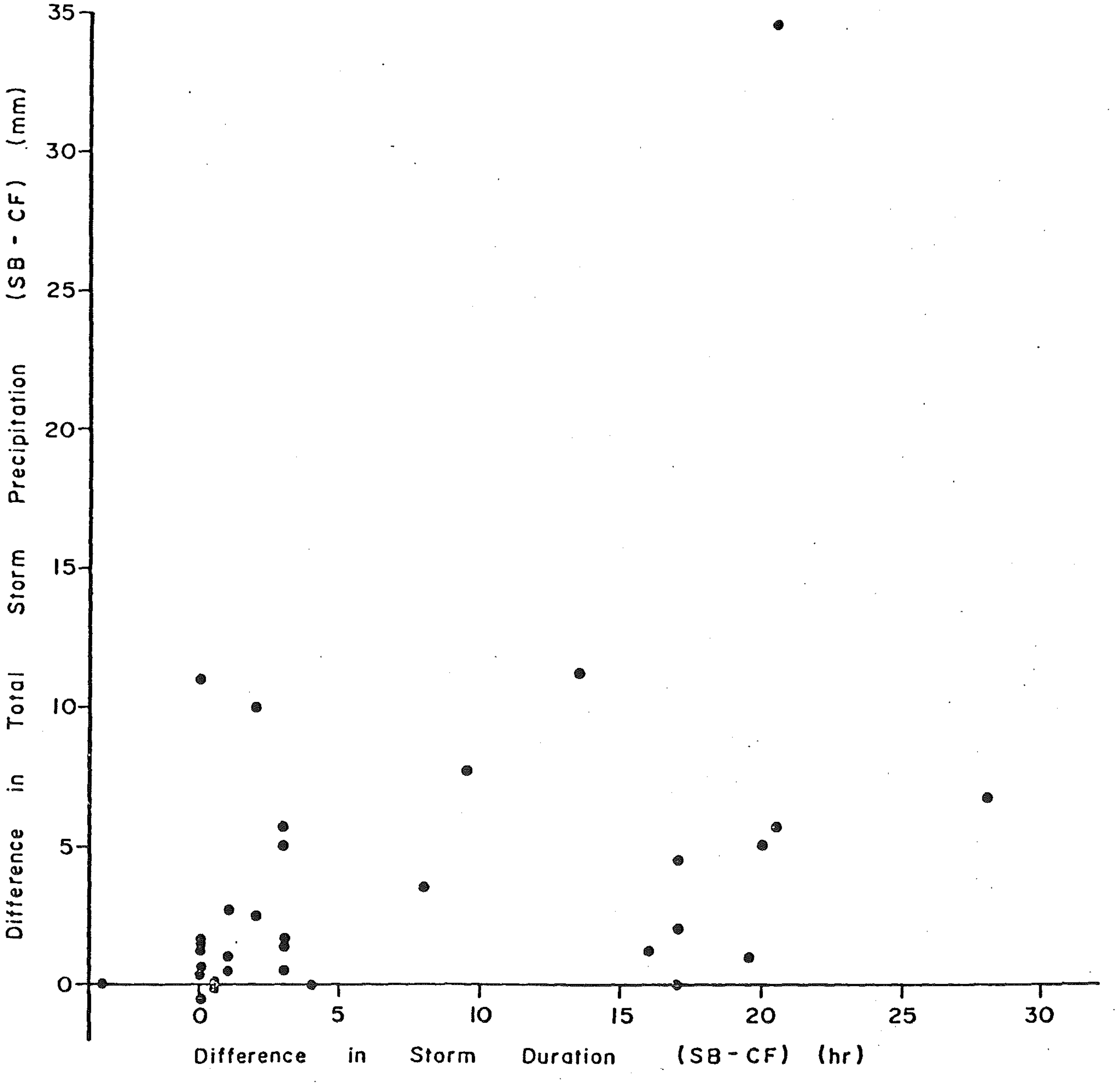
D.5


Difference in Total Storm Precipitation (SB-CF) (mm)
35
30
25
20
15
10
5
0
-0·8
-0·6
-0·4
-0·2
0
0·2
0·4
0·6
0·8
Difference in Storm Intensity (SB - CF) (mm.hr^{-1})

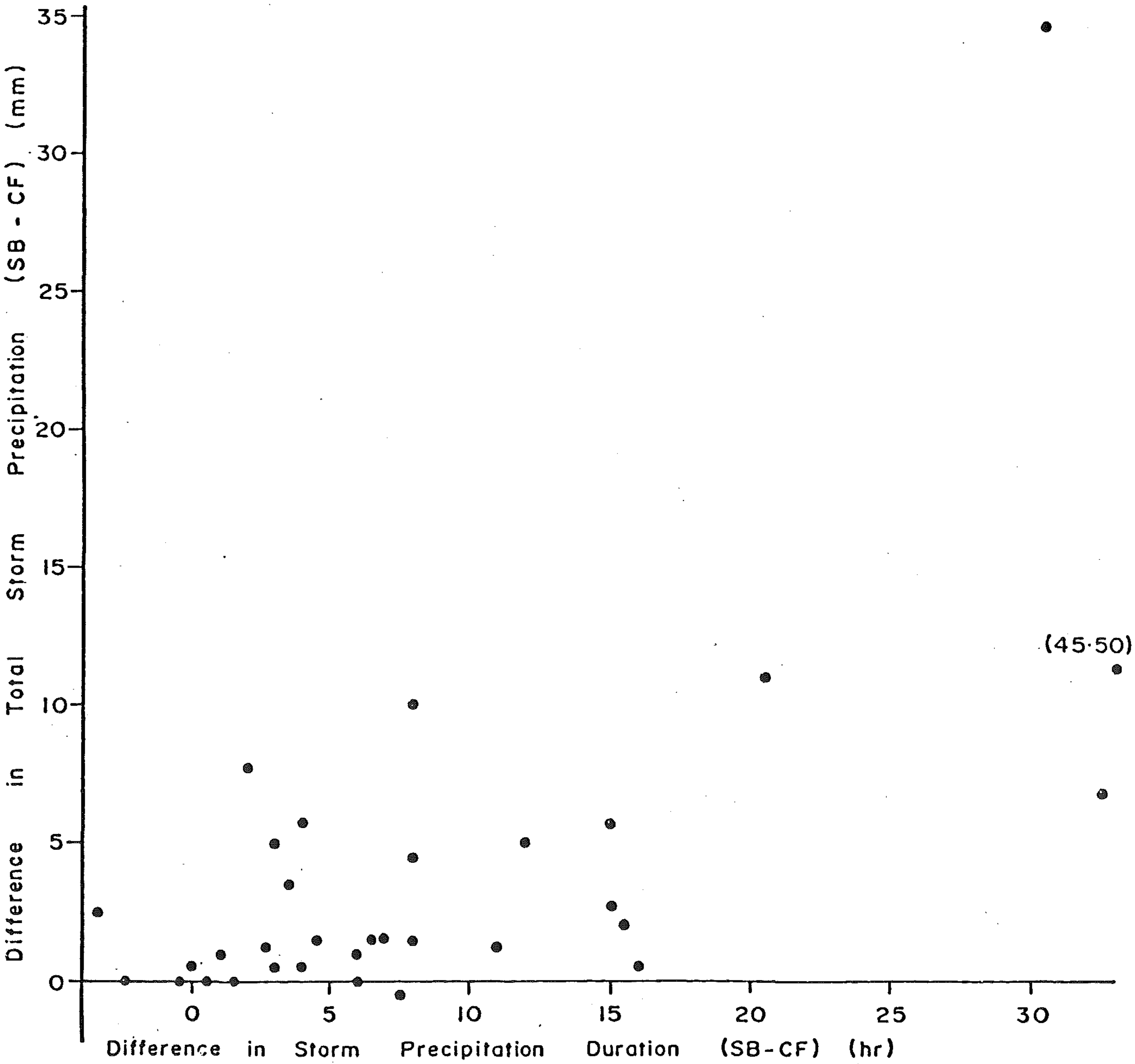
D.6


35
30
25
20
15
10
5
0
Difference in Total Storm Precipitation (SB - CF) (mm)
-0·6
-0·4
-0·2
0
0·2
0·4
0·6
0·8
1·0
Difference in Storm Precipitation Intensity (SB-CF) (mm.hr⁻¹)

D.7


35
30
25
20
15
10
5
0
0
5
10
15
20
25
30
Difference in Total Storm Precipitation (SB - CF) (mm)
Difference in Storm Duration (SB - CF) (hr)

D.8

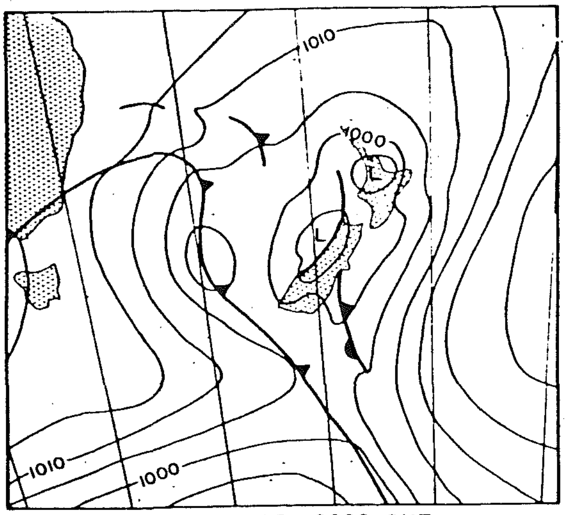

35
30
25
20
15
10
5
0
0
5
10
15
20
25
30
(45·50)
Difference in Total Storm Precipitation (SB - CF) (mm)
Difference in Storm Precipitation Duration (SB-CF) (hr)

APPENDIX E MAJOR SNOW PRODUCING STORMS.

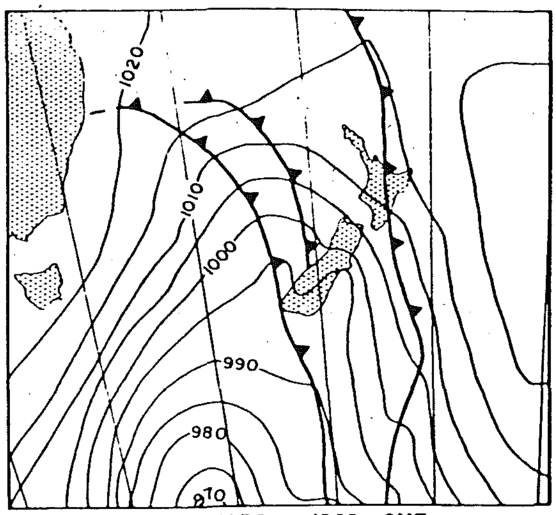
E.1. SURFACE SYNOPTIC CHARTS.

E.2. PRECIPITATION, WIND SPEEDS AND DIRECTIONS.

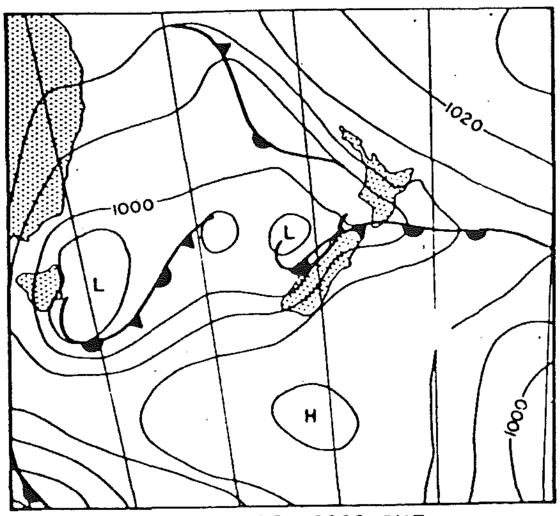
(Triangle on E.2 notes time of E.1 synoptic chart)



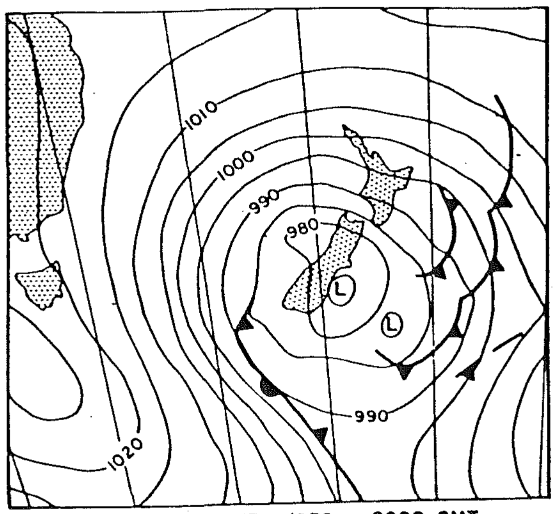
AUGUST 19 1975 0000 GMT



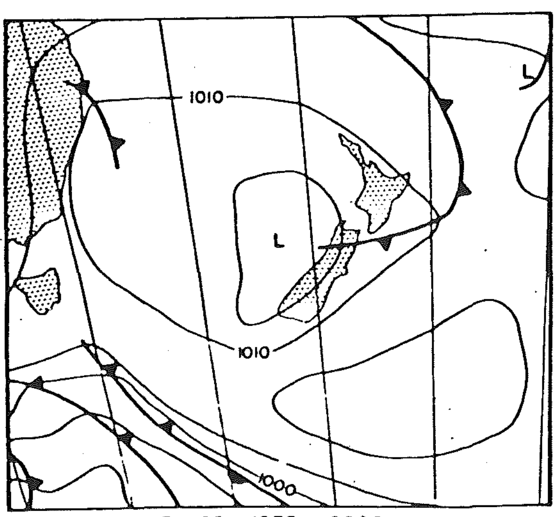
JUNE 9 1976 1200 GMT



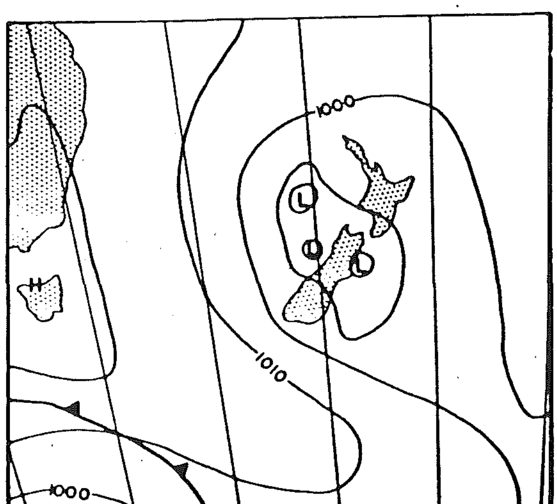
AUGUST 25 1975 0000 GMT



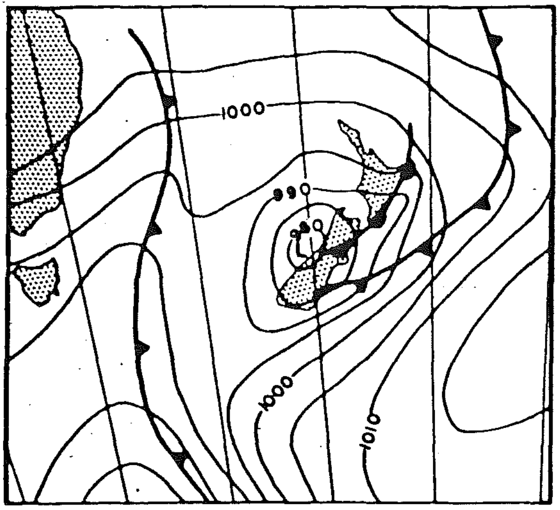
SEPTEMBER 13 1976 0000 GMT



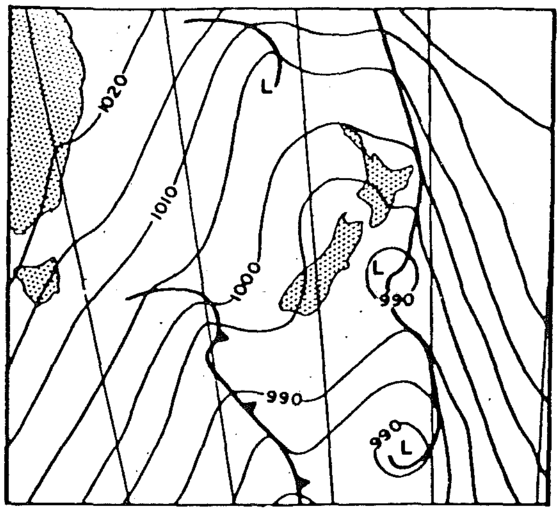
AUGUST 28 1975 0000 GMT



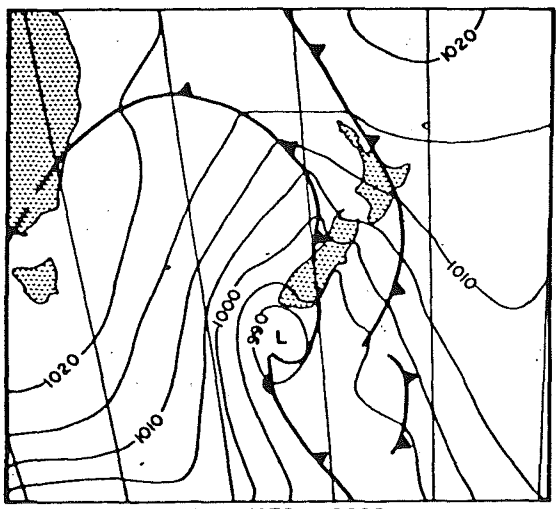
MAY 2 1977 0600 GMT



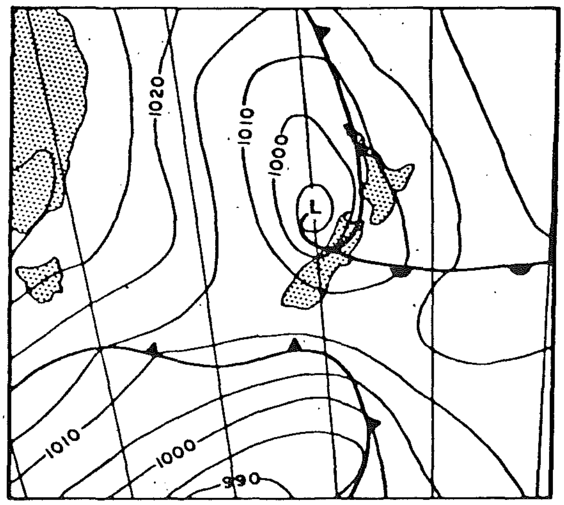
JUNE 28 1977 0600 GMT



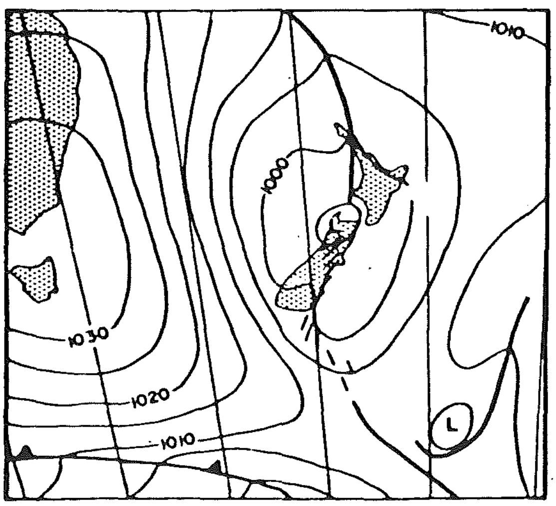
SEPTEMBER 15 1978 0000 GMT



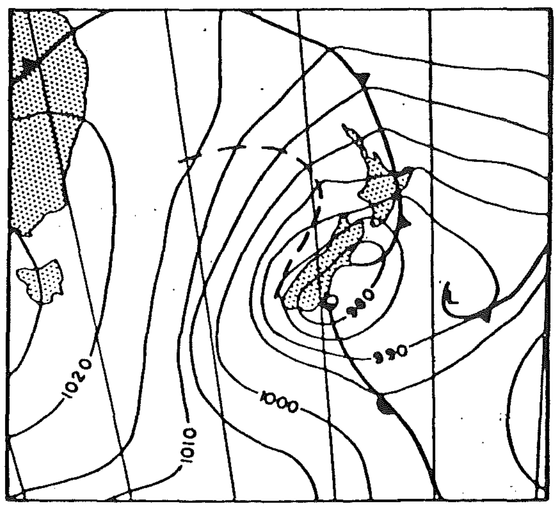
AUGUST 17 1978 0000 GMT



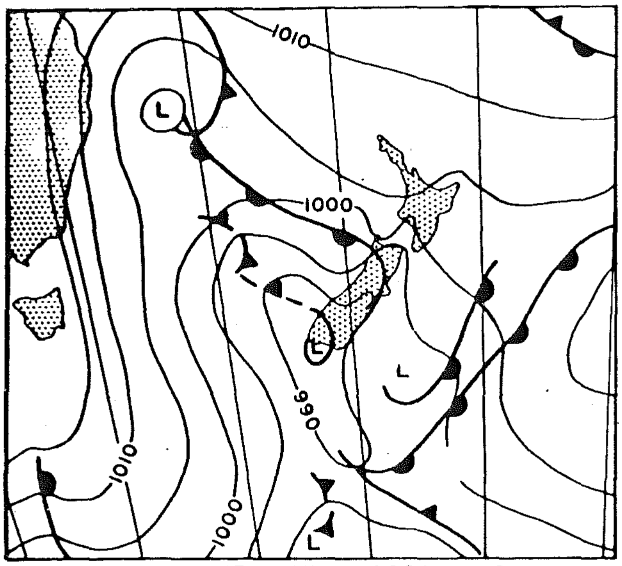
OCTOBER 19 1978 1200 GMT



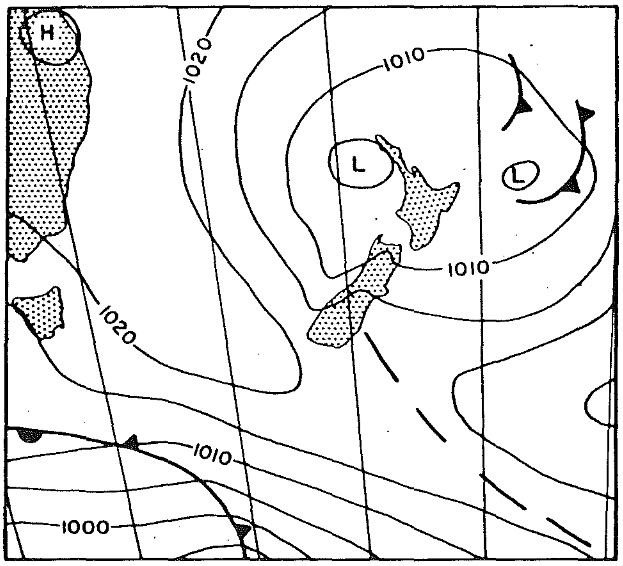
AUGUST 24 1978 0000 GMT



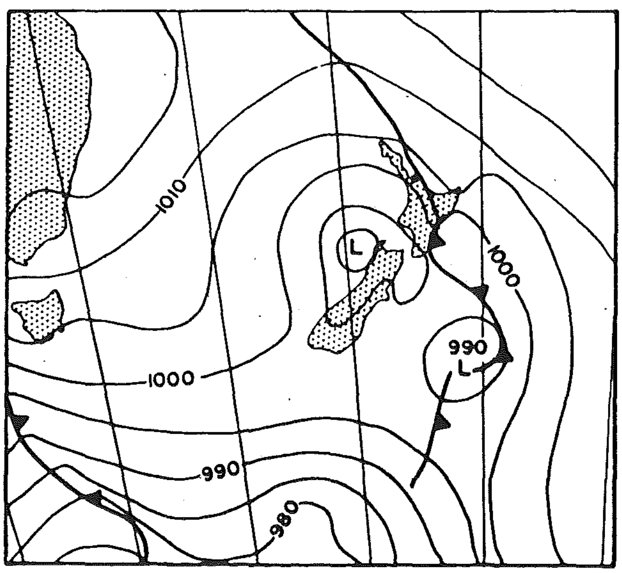
MAY 14 1979 0600 GMT



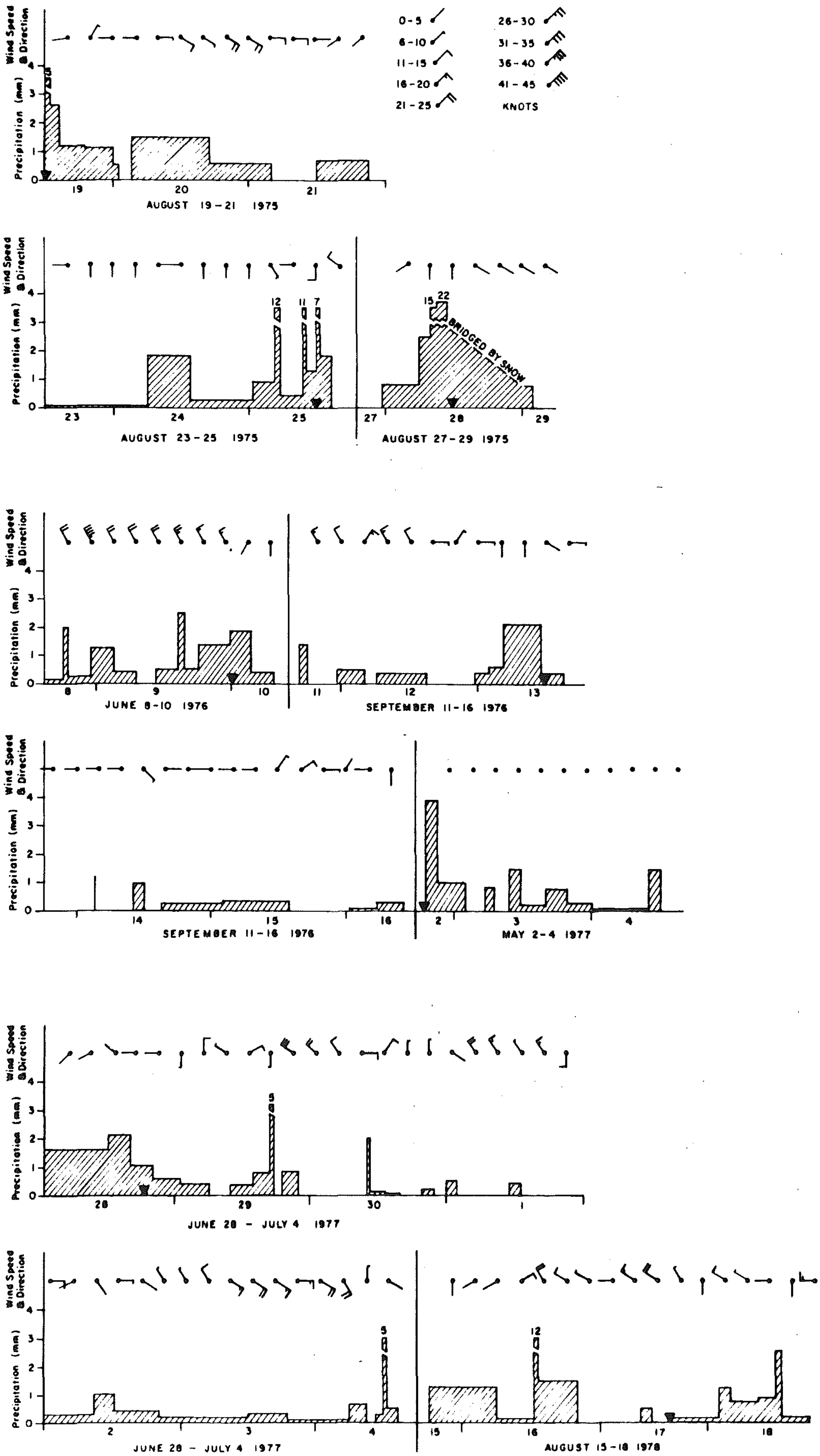
AUGUST 13 1979 0000 GMT



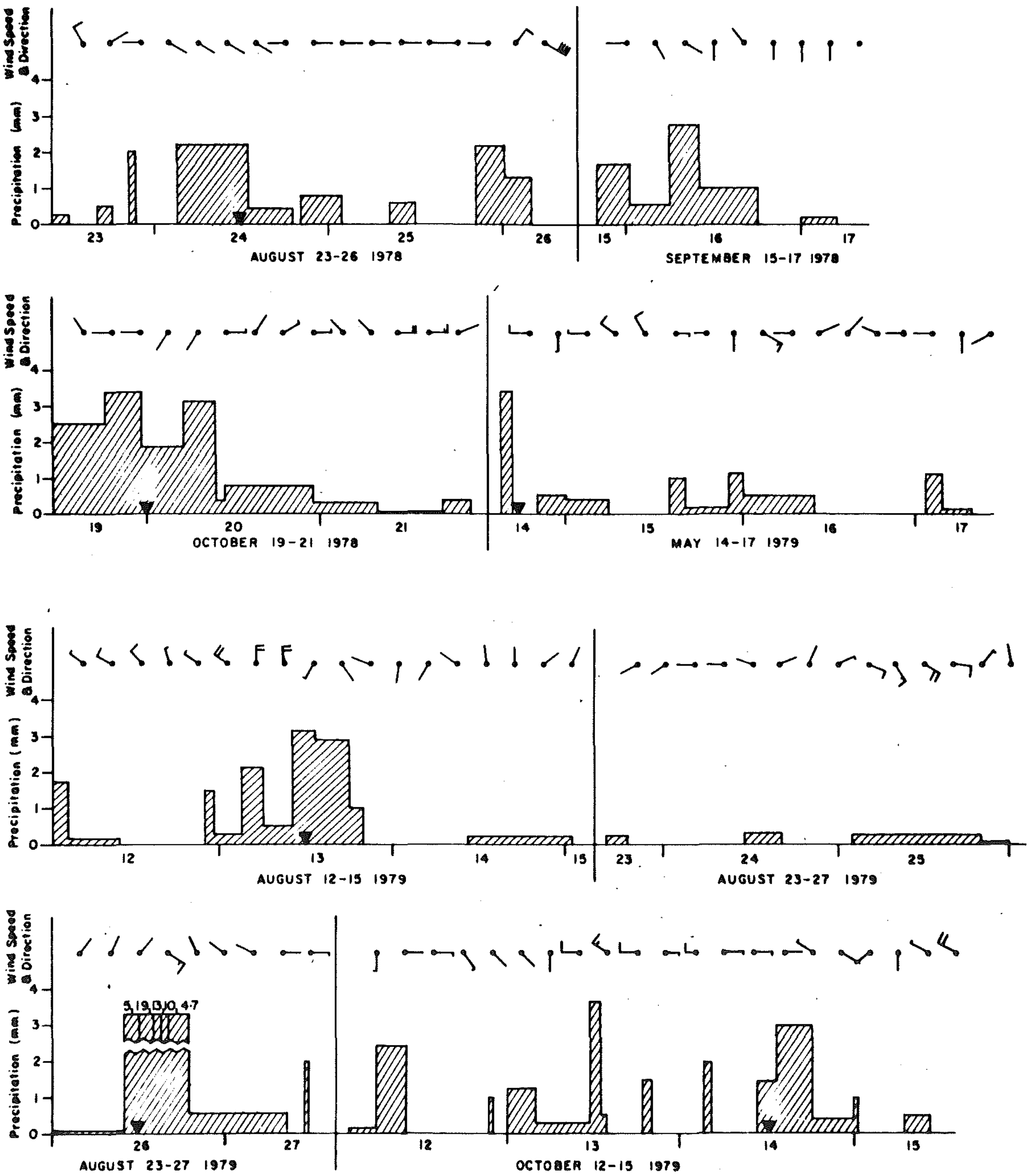
AUGUST 26 1979 0000 GMT



OCTOBER 14 1979 0000 GMT



Wind Speed & Direction
Precipitation (mm)
0-5
6-10
11-15
16-20
21-25
26-30
31-35
36-40
41-45
KNOTS
AUGUST 19-21 1975
AUGUST 23-25 1975
AUGUST 27-29 1975
BRIDGED BY SNOW
JUNE 8-10 1976
SEPTEMBER 11-16 1976
SEPTEMBER 11-16 1976
MAY 2-4 1977
JUNE 28 - JULY 4 1977
JUNE 28 - JULY 4 1977
AUGUST 15-18 1978



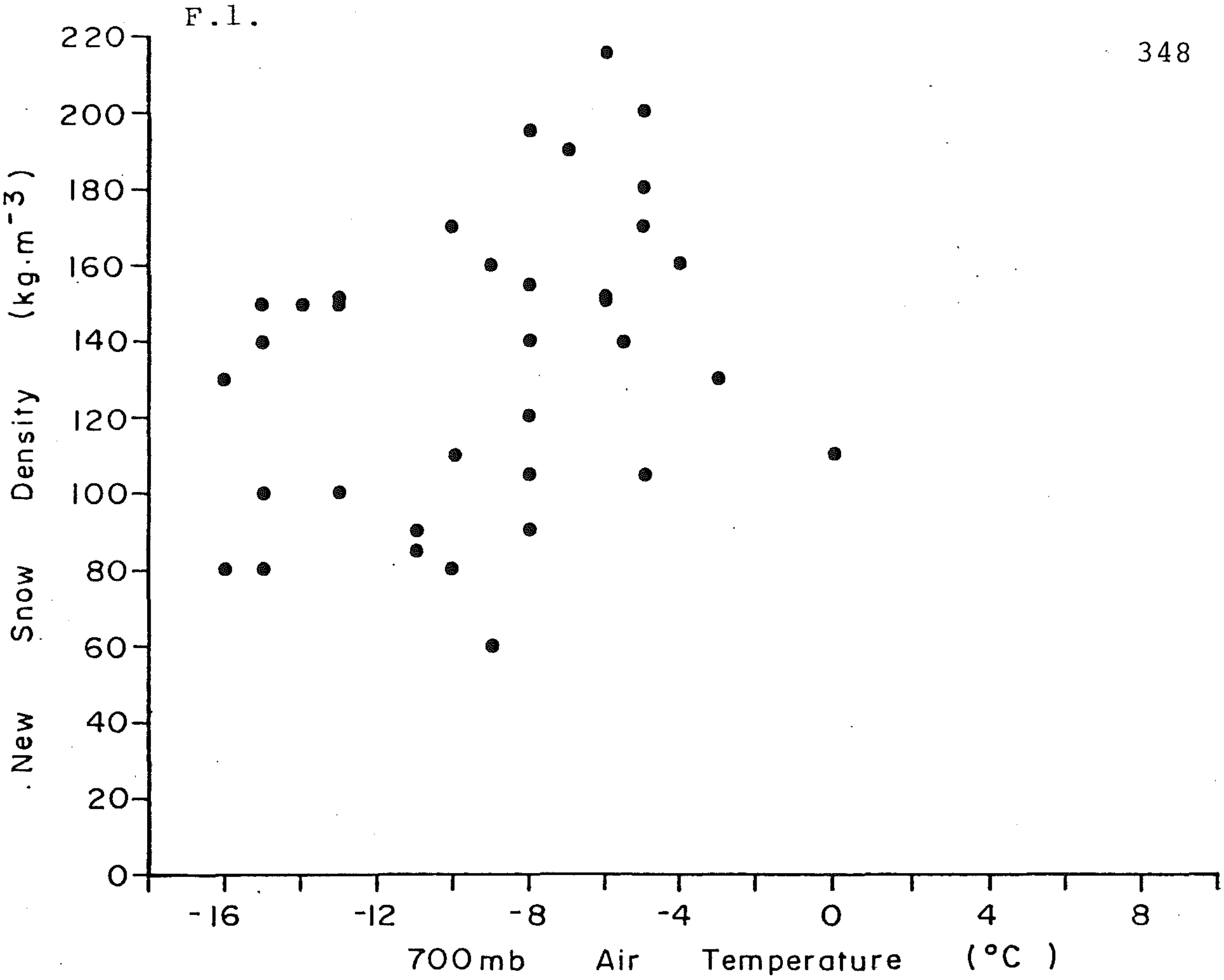
Wind Speed & Direction
Precipitation (mm)
0
1
2
3
4
23
24
25
26
AUGUST 23-26 1978
15
16
17
SEPTEMBER 15-17 1978
19
20
21
OCTOBER 19-21 1978
14
15
16
17
MAY 14-17 1979
12
13
14
15
AUGUST 12-15 1979
23
24
25
AUGUST 23-27 1979
5.1 9.3 13.0 4.7
26
27
AUGUST 23-27 1979
12
13
14
15
OCTOBER 12-15 1979

APPENDIX F RELATIONSHIP OF NEW SNOW DENSITY TO UPPER AIR TEMPERATURES.

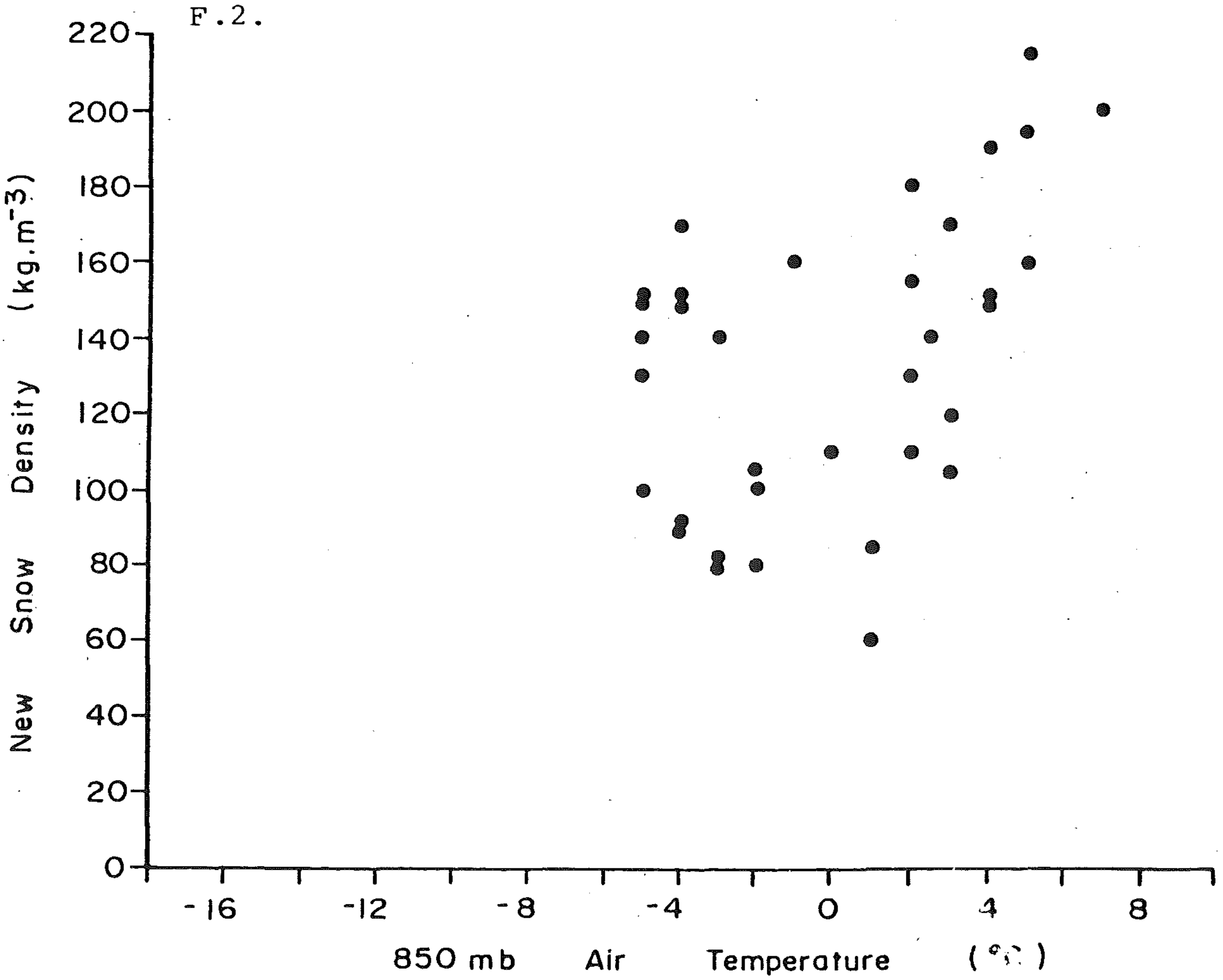
F.1. 700 mb AIR TEMPERATURE.

F.2. 850 mb AIR TEMPERATURE.

F.1.

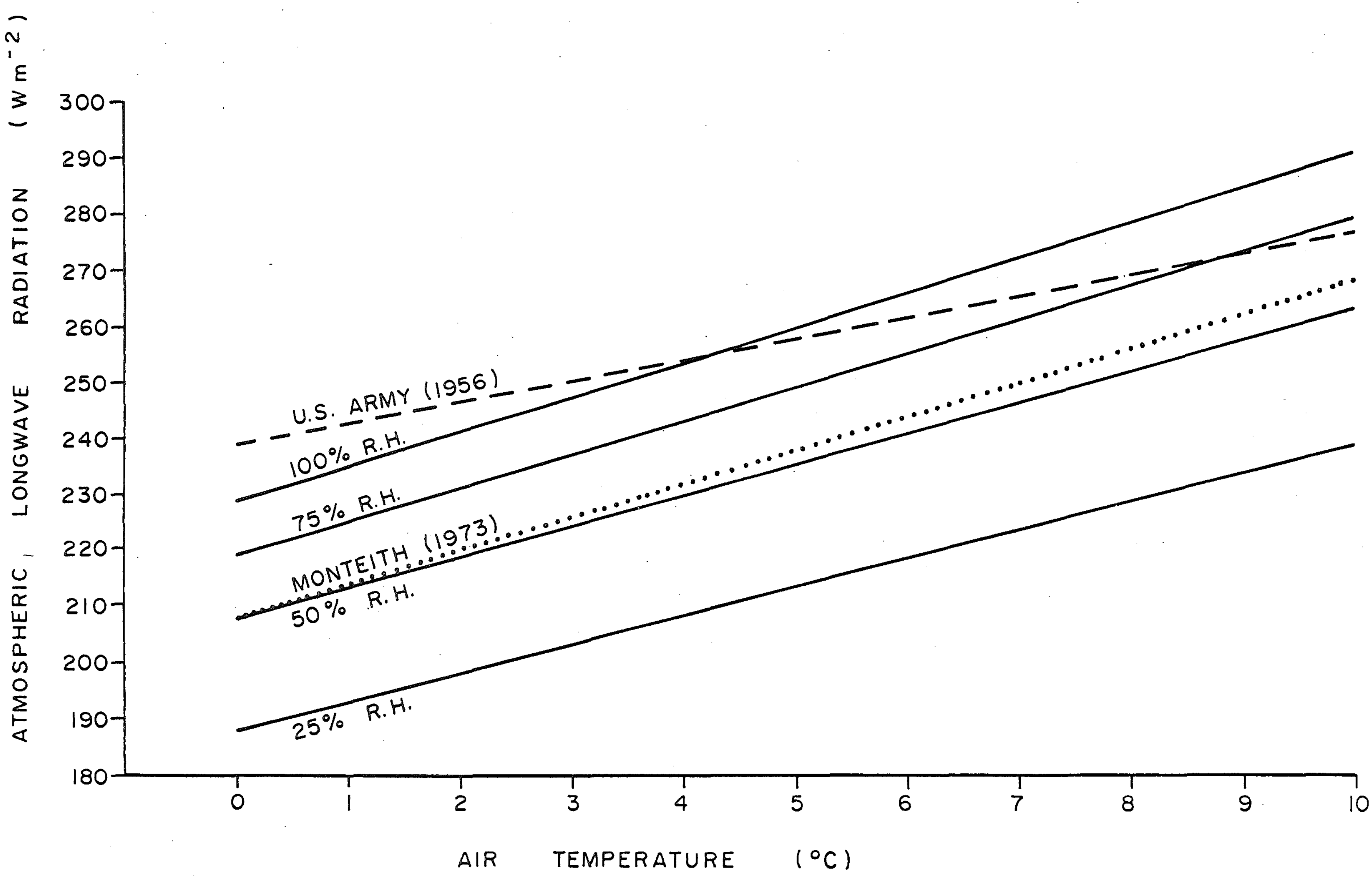

New Snow Density (kg.m^{-3})
220
200
180
160
140
120
100
80
60
40
20
0
-16
-12
-8
-4
0
4
8
700 mb Air Temperature (°C)

F.2.


New Snow Density (kg.m^{-3})
220
200
180
160
140
120
100
80
60
40
20
0
-16
-12
-8
-4
0
4
8
850 mb Air Temperature

APPENDIX G COMPARISON OF EQUATIONS FOR PREDICTING ATMOSPHERIC LONGWAVE RADIATION.

Brutsaert (1975) equation is represented by solid lines. RH refers to relative humidity values.



ATMOSPHERIC LONGWAVE RADIATION (W m⁻²)
300
290
280
270
260
250
240
230
220
210
200
190
180
U.S. ARMY (1956)
100% R.H.
75% R.H.
MONTEITH (1973)
50% R.H.
25% R.H.
0
1
2
3
4
5
6
7
8
9
10
AIR TEMPERATURE (°C)

APPENDIX H CLOUD ADJUSTMENT FACTORS FOR PREDICTING ATMOSPHERIC RADIATION.

CLOUD TYPE	TYPICAL CLOUD HEIGHT (km)	COEFFICIENT Ω
CIRRUS	12.20	0.04
CIRROSTRATUS	8.39	0.08
ALTOCUMULUS	3.66	0.17
ALTOSTRATUS	2.14	0.20
STRATOCUMULUS	1.22	0.22
STRATUS	0.46	0.24
FOG	0.00	0.25

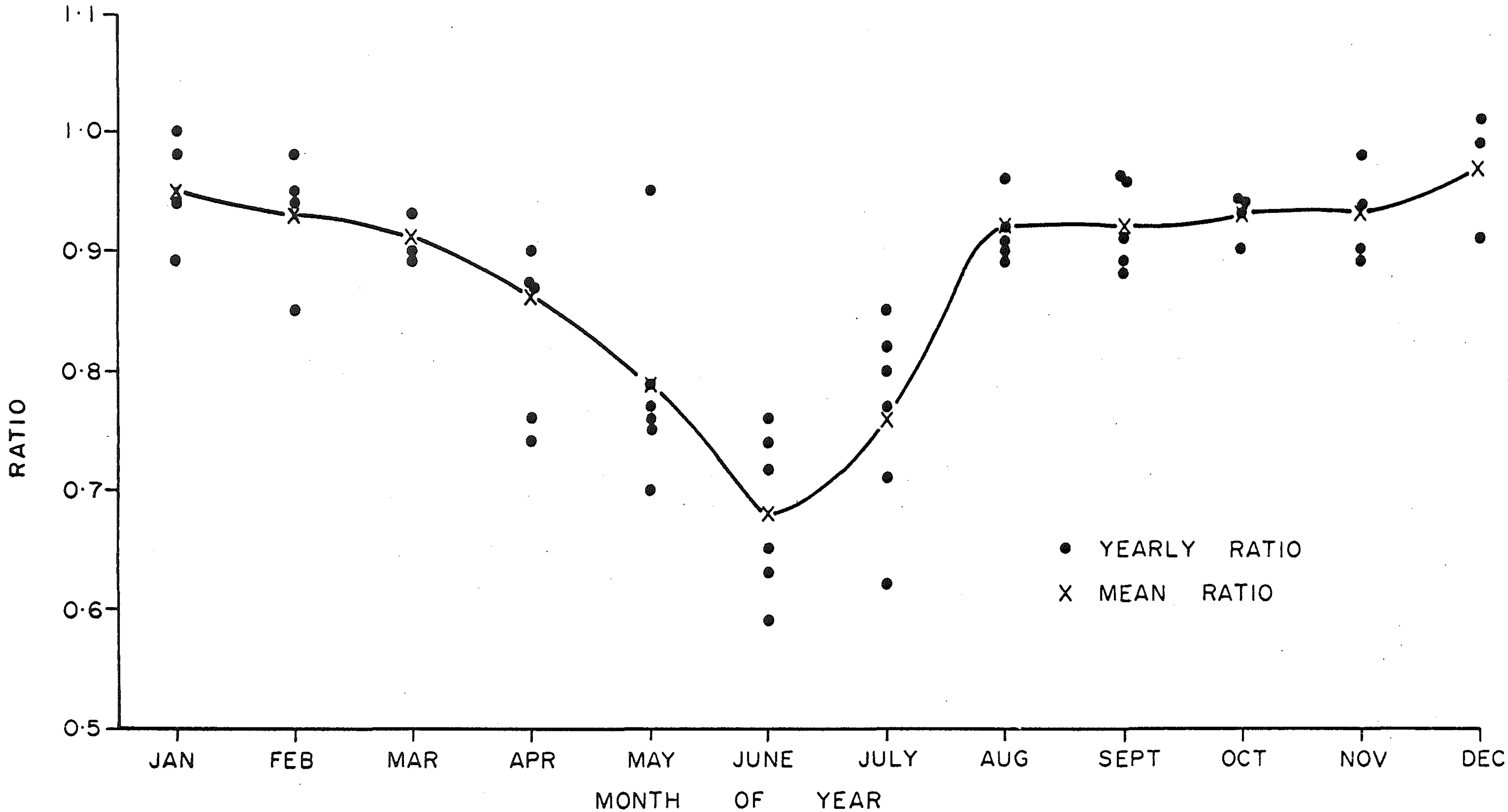
Values of the coefficient Ω used to allow for decreasing cloud temperatures with height in equation (modified after Sellers, 1965).

APPENDIX I RELATIONSHIP OF SHORTWAVE RADIATION BETWEEN THE SB and CF CLIMATE STATIONS.

	1974			1973			1972			1971			1970			1969		
	C.F.	S.B.	RATIO	C.F.	S.B.	RATIO	C.F.	S.B.	RATIO	C.F.	S.B.	RATIO	C.F.	S.B.	RATIO	C.F.	S.B.	RATIO
JANUARY	21.86	21.82	1.00	24.37	22.95	0.94	22.19	21.65	0.98				22.70	20.23	0.89			
FEBRUARY	17.00	14.45	0.85	20.85	19.68	0.94	19.93	18.97	0.95				21.15	20.77	0.98			
MARCH	13.36	12.02	0.90	14.99	13.27	0.89	14.57	13.48	0.93				12.60	11.47	0.91			
APRIL	8.21	6.24	0.76	10.93	9.46	0.87	9.55	9.00	0.94	11.52	9.72	0.84	11.73	10.51	0.90	11.43	9.88	0.87
MAY	6.32	4.40	0.70	5.99	4.48	0.75	6.70	5.32	0.79	6.53	4.98	0.76	7.79	6.03	0.77	6.11	5.78	0.95
JUNE	5.57	3.31	0.59	5.36	3.39	0.63	5.78	4.19	0.72	(4.94)	3.22	(0.65)	5.40	4.02	0.74	6.03	4.56	0.76
JULY	5.57	3.43	0.62	6.07	4.31	0.71	5.70	4.40	0.77	(5.44)	(4.48)	(0.82)	5.90	5.03	0.85	6.41	5.15	0.80
AUGUST				7.45	6.83	0.92	10.13	9.17	0.90	8.58	7.79	0.91	7.91	7.58	0.96	10.64	9.42	0.89
SEPTEMBER				14.32	12.65	0.88	13.02	12.56	0.96	12.73	12.94	1.02	11.93			12.73	11.64	0.91
OCTOBER				17.55	16.54	0.94	17.96	16.67	0.93	18.63	16.75	0.90	17.92			20.10	18.97	0.94
NOVEMBER				20.52	18.34	0.89	21.36	19.30	0.90	20.18	18.89	0.94	23.79			24.20	23.62	0.98
DECEMBER				19.93	20.18	1.01	22.61	22.28	0.99	21.98	20.06	0.91	24.37			23.32	17.21	0.74

APPENDIX

MEAN MONTHLY SHORTWAVE RADIATION M J d^{-1}
FROM CRAIGIEBURN FOREST AND SKI BASIN
METEOROLOGICAL STATIONS.
RATIO = S.B./C.F.


RATIO
1·1
1·0
0·9
0·8
0·7
0·6
0·5
JAN
FEB
MAR
APR
MAY
JUNE
JULY
AUG
SEPT
OCT
NOV
DEC
MONTH OF YEAR
● YEARLY RATIO
X MEAN RATIO

APPENDIX J WEEKLY DEGREE HOURS FOR AIR, SNOW AND GROUND TEMPERATURES, JUNE TO NOVEMBER 1979.

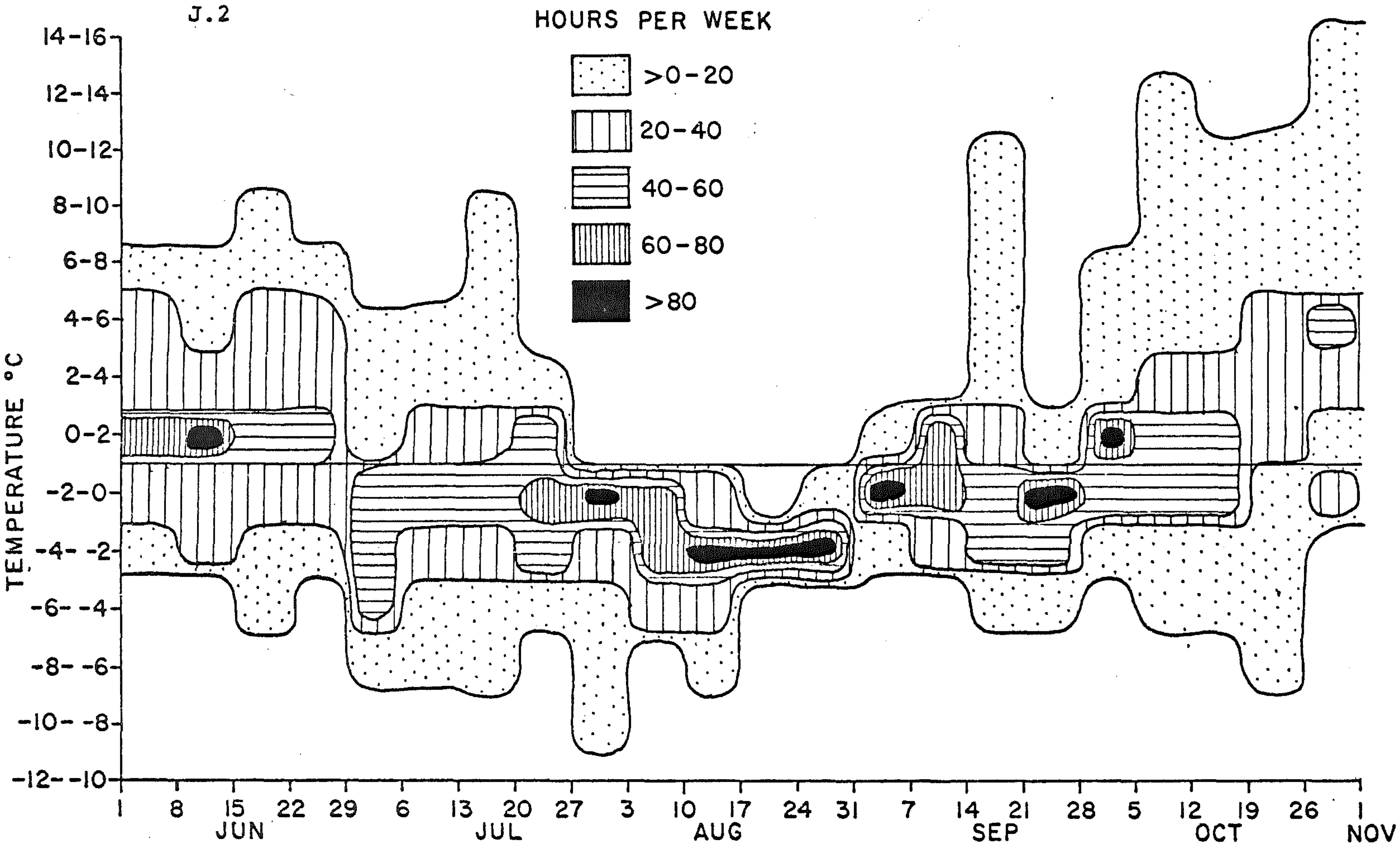
J.1. +1500 mm THERMISTOR RECORD

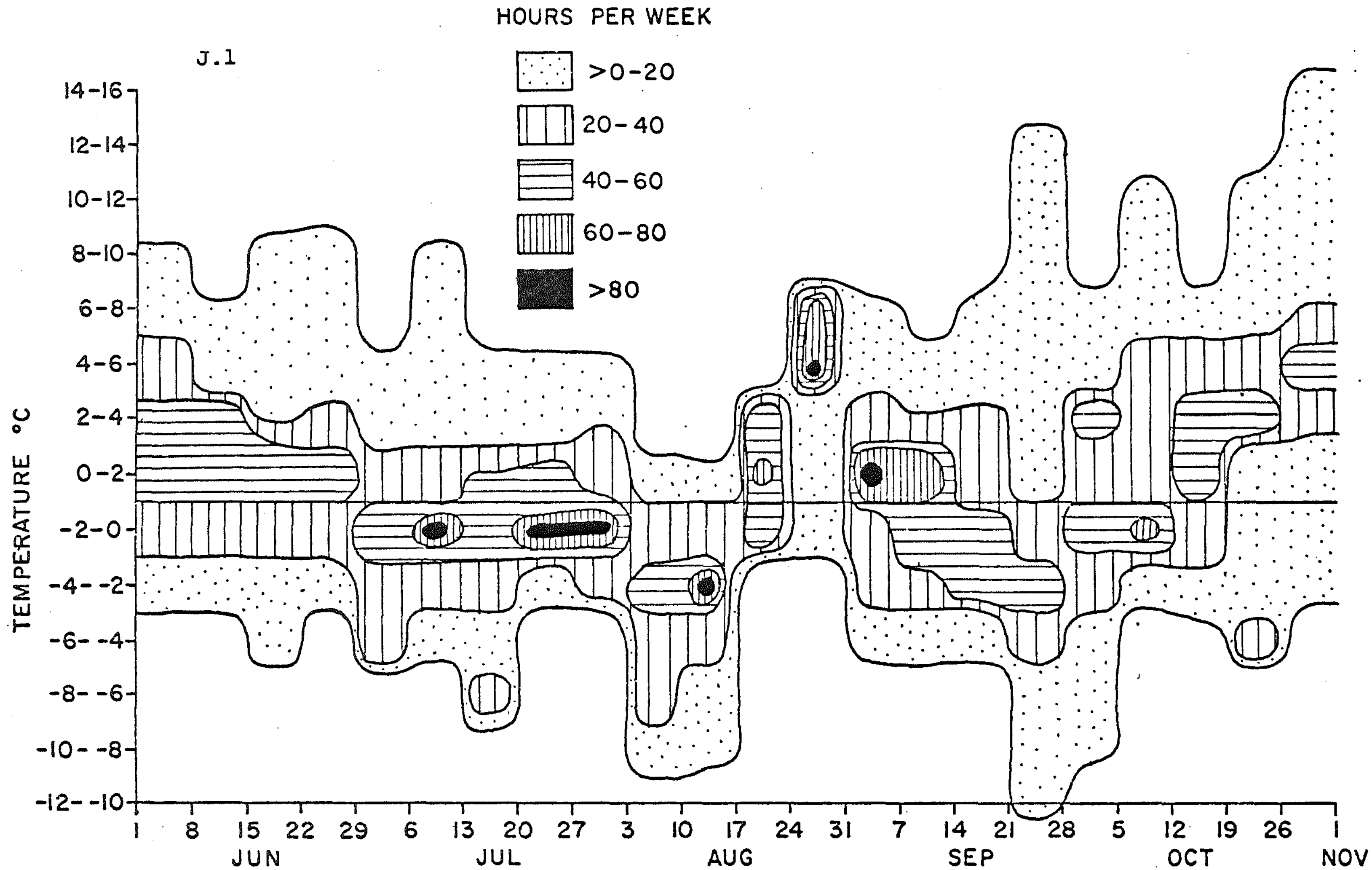
J.2. +300 mm THERMISTOR RECORD

J.3. +50 mm THERMISTOR RECORD

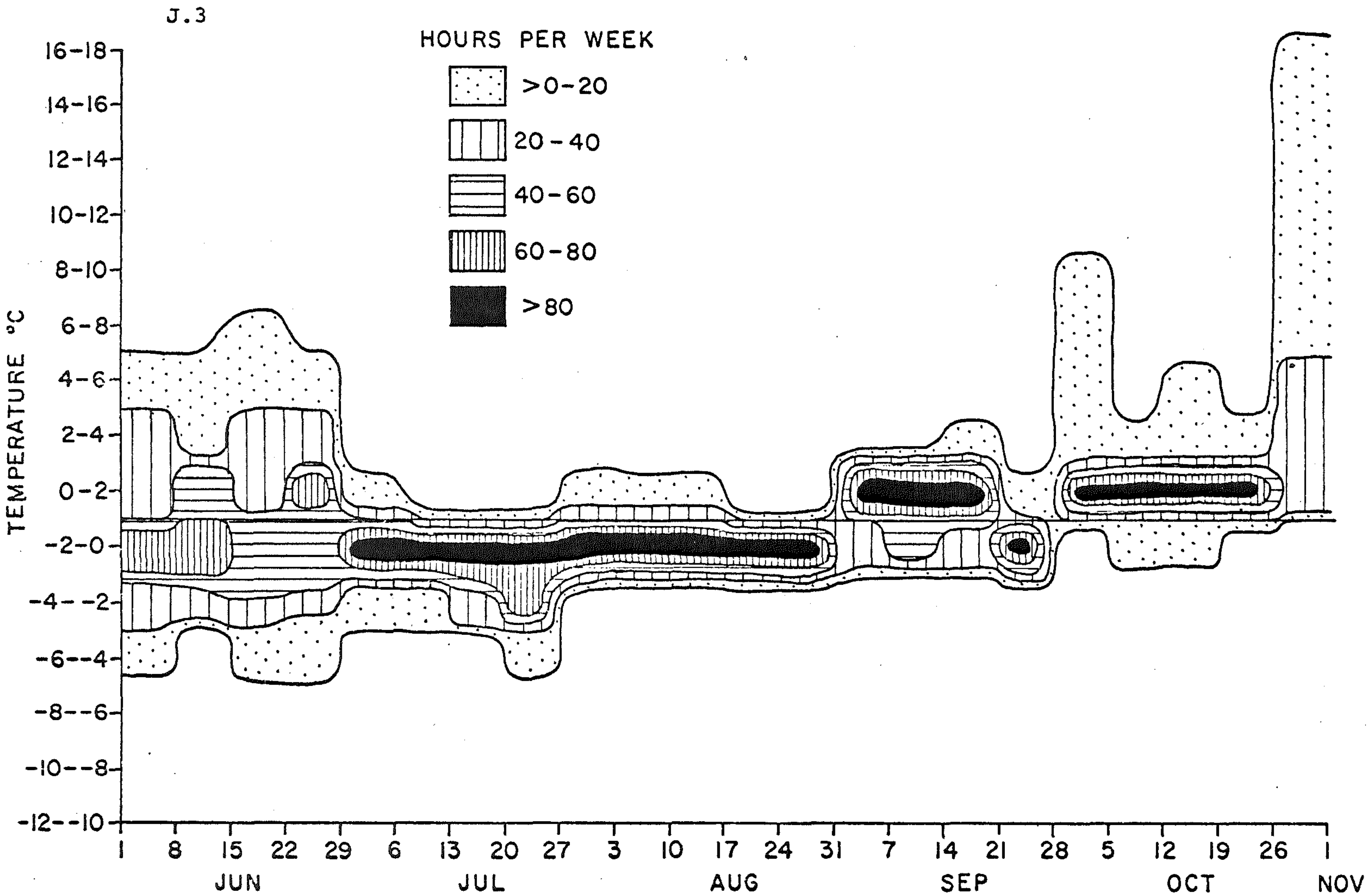
J.4. -50 mm THERMISTOR RECORD

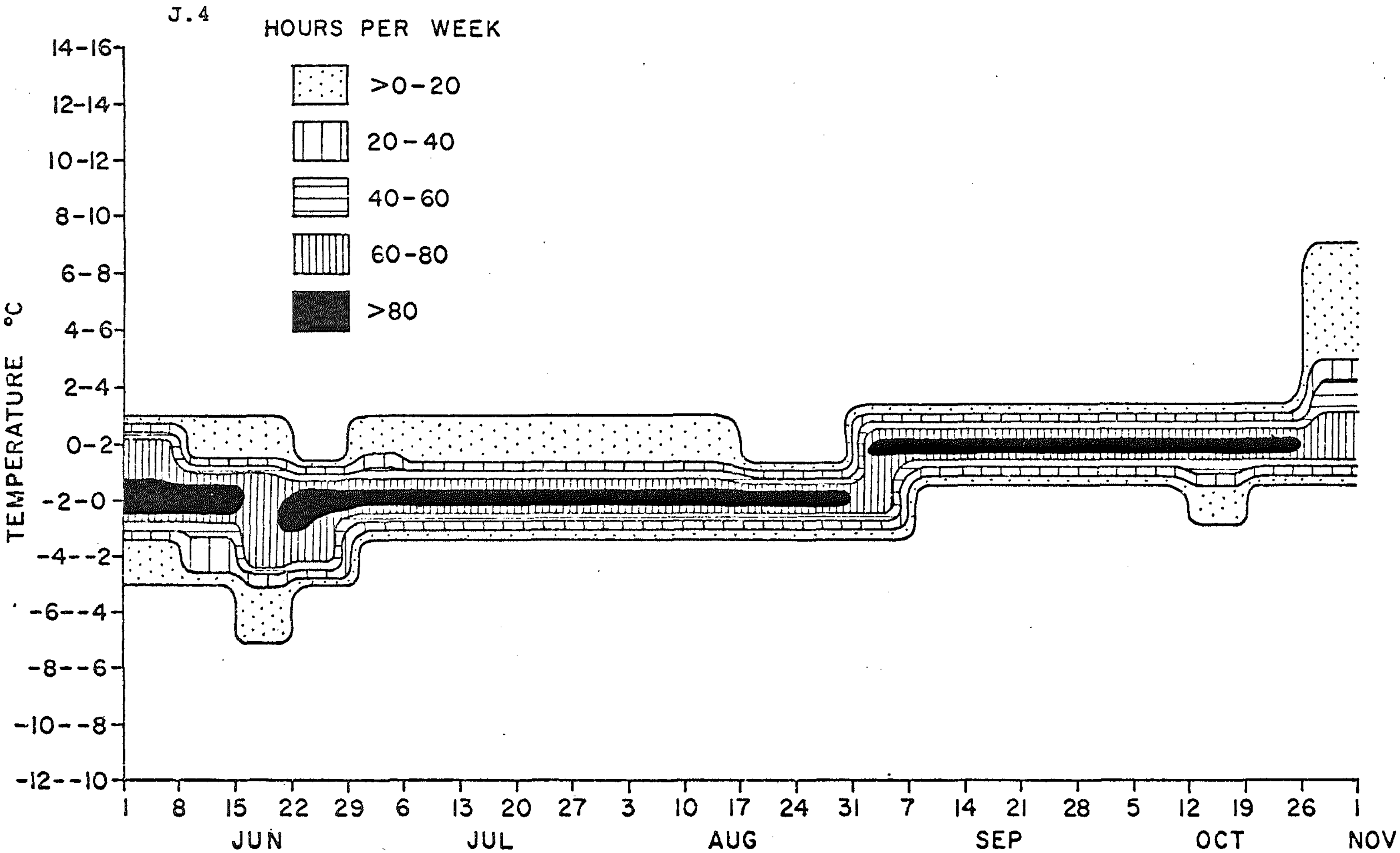
All records are from the #6 thermistor station at an elevation of 1690 m on Mt Cockayne. Hours per week denote accumulated hours per week at the specified temperature range.


J.2
HOURS PER WEEK
>0-20
20-40
40-60
60-80
>80
TEMPERATURE °C
14-16
12-14
10-12
8-10
6-8
4-6
2-4
0-2
-2-0
-4- -2
-6- -4
-8- -6
-10- -8
-12- -10
1 8 15 22 29 6 13 20 27 3 10 17 24 31 7 14 21 28 5 12 19 26 1
JUN
JUL
AUG
SEP
OCT
NOV



HOURS PER WEEK
>0-20
20-40
40-60
60-80
>80
J.1
TEMPERATURE °C
14-16
12-14
10-12
8-10
6-8
4-6
2-4
0-2
-2-0
-4--2
-6--4
-8--6
-10--8
-12--10
1 8 15 22 29 6 13 20 27 3 10 17 24 31 7 14 21 28 5 12 19 26 1
JUN JUL AUG SEP OCT NOV


J.3
HOURS PER WEEK
>0-20
20-40
40-60
60-80
>80
TEMPERATURE °C
16-18
14-16
12-14
10-12
8-10
6-8
4-6
2-4
0-2
-2-0
-4--2
-6--4
-8--6
-10--8
-12--10
1
8
15
22
29
6
13
20
27
3
10
17
24
31
7
14
21
28
5
12
19
26
1
JUN
JUL
AUG
SEP
OCT
NOV


J.4
HOURS PER WEEK
>0-20
20-40
40-60
60-80
>80
TEMPERATURE °C
14-16
12-14
10-12
8-10
6-8
4-6
2-4
0-2
-2-0
-4--2
-6--4
-8--6
-10--8
-12--10
1 8 15 22 29 6 13 20 27 3 10 17 24 31 7 14 21 28 5 12 19 26 1
JUN
JUL
AUG
SEP
OCT
NOV